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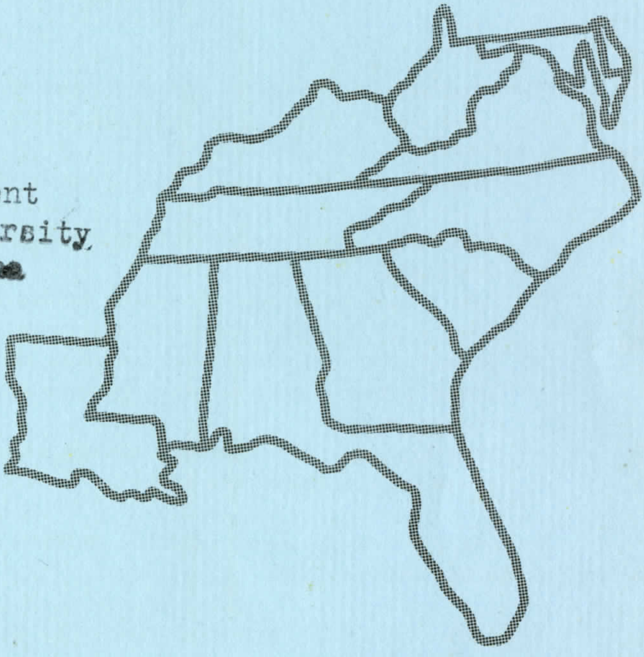

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SOUTHEASTERN GEOLOGY

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Vol. 14, No. 4

1972

GIBBSITE IN WEATHERED GRANITIC ROCKS OF ALABAMA[1/]

By

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ABSTRACT

Gibbsite, which is a product of humid tropical weathering, occurs locally in saprolite of granitic rocks in Alabama, although the climate today is more temperate than tropical. The gibbsite may be a relic mineral of past weathering in a climate that was warmer and more humid than the current climate.

Gibbsite in the saprolite reflects the mineralogy of the parent rock. Plagioclase feldspar has altered to gibbsite; whereas, orthoclase feldspar, subjected to the same weathering, has altered to kaolinite. This is in accordance with the weathering sequence noted in humid tropical zones such as Guyana. Biotite has altered to kaolinite, but field relations indicate that it may have passed through an intermediate clay-mica (illite) and interlayered illite-vermiculite stage.

The occurrence of gibbsite is related to topography; it occurs in the upland soils of the Ashland and Opelika Plateaus, and it does not occur in substantial amounts of soils on valleys and terraces of Pleistocene and Holocene age. The uplands containing gibbsite are well drained, and silica leached from quartz and other silicates has been removed from the system; otherwise the gibbsite is resilicated to kaolinite. The weathering sequence in poorly drained areas is both plagioclase and orthoclase altering to kaolinite.

INTRODUCTION

Gibbsite, a common constituent of weathered rock in humid tropical areas (Harder, 1952), occurs locally in mature soils derived from igneous and metamorphic rocks in the southeastern United States (McCracken, 1959). Although this region is hot during the summer months, the climate cannot be described as humid tropical, because the annual rainfall is less than 2 meters and the winter months are cool. Gibbsite occurs with quartz in the saprolite although the two minerals

1/ Publication authorized by the State Geologist, Geological Survey of Alabama.

are not mutually stable in a closed system (Kittrick, 1969).

Saprolite containing gibbsite and kaolinite has been investigated as potential alumina reserves in the southeastern United States (Patterson, 1967). One of the problems encountered was the erratic distribution of gibbsite in the weathered rock.

Considerable work on soil mineralogy has been performed by the agronomy departments of agricultural colleges in the Southeast. Hendricks and Fry (1930) classified some soils in Georgia and North Carolina as "bauxite" because the clay fractions contained more alumina than silica. McCracken (1959), in describing physical and chemical properties of soil samples collected over granitic rocks in Virginia, North Carolina, and Alabama, showed that the clay fractions are composed predominantly of gibbsite, kaolinite, and interlayered vermiculite. Bryant and Dixon (1964) described the clay fraction of Red-Yellow Podsolic soils derived from quartz-mica schist in Alabama, as consisting of kaolinite and gibbsite.

Fieldwork and Methods of Investigation

The saprolite study was conducted as part of a bauxite investigation in Alabama, because the weathered rocks of the Piedmont physiographic province are the ultimate source of most high-alumina deposits in the Coastal Plain sediments (Clarke, 1966, p. 913). This report is based, in part, on previous investigations to evaluate saprolite as a source of alumina in an area that extends from Alabama to northern Virginia. The mineral data reported here was collected between 1967 and 1971. Mineral identifications were made by X-ray diffraction supported by chemical analyses (Clarke, 1971, p. 79, 80). Samples collected were correlated with position in the soil profile, rock type, mineralogy of the host rock, and topography.

Climate and Physiography

Alabama is in the warm temperate zone. The mean annual rainfall of the Piedmont province ranges from 115 to 150 centimeters and averages about 132 centimeters (U. S. Weather Bureau, 1968). The heaviest rainfall occurs during the winter, early spring, and midsummer. Winters are mild, the temperatures averaging about 9.5°C; the summers are warm, the temperatures averaging about 27°C.

Adams and others (1926, p. 26-27) divided the Piedmont uplands into the Ashland Plateau and the Opelika Plateau (Figure 1). The surface of the Ashland Plateau lies at altitudes of 335 to 365 meters above msl (mean sea level) in Cleburne County and 150 to 180 meters above msl along the southwestern border where it merges with the Coastal Plain in Chilton and Elmore Counties. The plateau is deeply dissected by stream erosion, and in many areas the old surface has been completely removed. In other regions remnants of the plateau surface

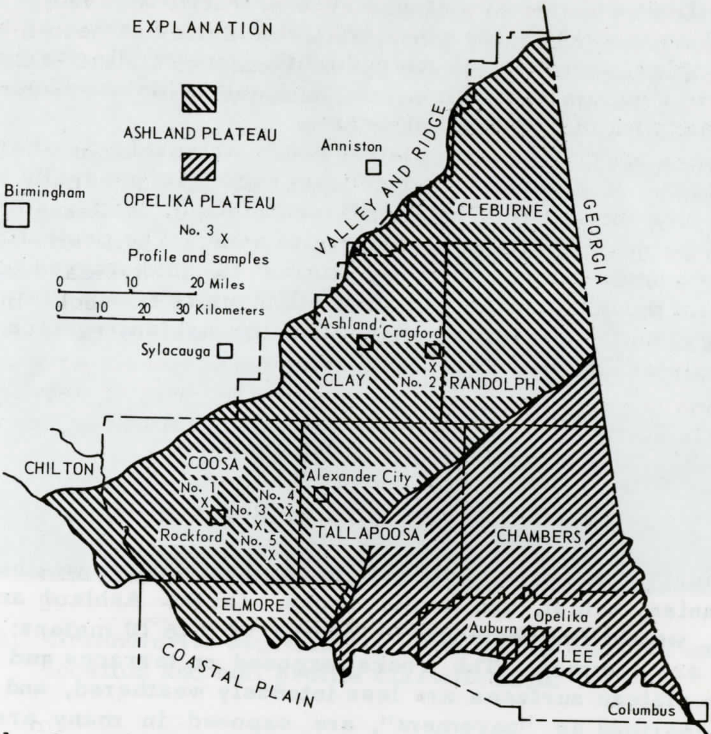

Figure 1. Location of study area.

occur as long, irregular, flat-topped uplands.

The Opelika Plateau lies at altitudes ranging from about 240 meters above msl in the northern part of Chambers County to about 180 meters above msl along the Coastal Plain in Lee County. Like the Ashland Plateau, the Opelika Plateau is dissected by stream erosion, but it has not been eroded to the same degree. Large areas of the original plateau remain as undulating uplands.

Geologic Setting

The rocks underlying the Piedmont physiographic province are predominantly phyllite, schist, and gneiss that have resulted from regional metamorphism of clastic sediments. Both silicic and mafic intrusives occur in the Piedmont region of Alabama.

Granitic rocks in Alabama are classified here into two generalized types on the basis of structure and petrography. One type designated as massive granitic rock has granitic texture and sharp contacts with adjacent rocks. Plagioclase feldspars are more abundant in the granitic rocks of Alabama than the orthoclase feldspars, and most exposures are classified as granodiorites and quartz diorites rather than true granites (Gault, 1945). The other type, designated here as a granitic gneiss complex, consists of granodiorite, quartz diorite and biotite gneiss that is generally gneissic and includes schistose bands that are conformable with the regional structure.

Rocks underlying the plateaus are intensely weathered. The soft weathered rock horizon, termed saprolite, is generally between 5 and 20 meters thick, but "ribs" or masses of hard, less weathered rock commonly project upward in the saprolite zone. The projecting "ribs" commonly contain cores of slightly altered to unweathered rock. The saprolite on the Ashland and Opelika Plateaus is similar in thickness, mineralogy, and appearance, except that the Ashland surface has been more deeply eroded by intersecting streams.

RESULTS AND DISCUSSION

Massive Granitic Rocks

Massive granitic rocks are more resistant to weathering than biotite gneiss, schist, and mafic rocks. On the Ashland and Opelika Plateaus, weathering extends to depths of 2 to 10 meters; hard rock outcrops are common. The rocks exposed on terraces and in valleys below the plateau surfaces are less intensely weathered, and hard outcrops, described as "pavement", are exposed in many areas. Two granodiorite outcrops investigated in detail are discussed here as typical examples.

One exposure, located about 2.7 kilometers west of Rockford in Coosa County (location 1 on Figure 1) on the Ashland Plateau, includes both soft saprolite and a hard outcrop of weathered granodiorite with unweathered rock exposed in the ditch. The plateau surface in this region is an undulating upland ranging from about 220 meters to a maximum of 250 meters above msl. Sample profile no. 1 was collected from a low rise at an altitude of about 230 meters above msl. The location of samples collected within the saprolite profile is shown in Figure 2. The description and mineralogy of the profile is given in Table 1.

The argillaceous (B soil) horizon contains more kaolinite than gibbsite, but the saprolite below (C1 soil) horizon contains more gibbsite than kaolinite. The hard "rib" of partly weathered rock contains orthoclase as a major constituent and only traces of kaolinite, whereas, the soft saprolite contains kaolinite as a major constituent and only a trace of orthoclase. Gibbsite is the most abundant constituent of both samples. There is no indication of any difference in mineralogy of the parent or source rock.

The second granodiorite sample location (location 2, Figure 1) is an outcrop near the western edge of the Cragford arsenic-lead-gold

Figure 2. Granodiorite saprolite location no. 1, showing sample location and soil profile classification.

district. It is in a valley at altitude 220 meters above msl, which is about 80 meters below the Ashland Plateau. The clay fraction is composed of kaolinite with minor amounts of clay mica (illite) and no feldspars. The absence of feldspars indicates complete alteration of the original rock.

The main difference in the clay mineralogy of the two profiles is that gibbsite is a major component of the clay (size) fraction of the saprolite samples collected from the Ashland Plateau, but it is not identified in samples collected in the valley. The outcrop is located in a valley less than 5 meters above the present stream bed and subterranean drainage has been restricted at various times in the past. The original mineral composition of both rock samples is very similar in that the rocks are composed of predominantly plagioclase and quartz with a minor amount of orthoclase and biotite.

Granitic Gneiss Complex

A granitic gneiss complex exposed on the uplands of both the Ashland and Opelika Plateaus is covered by a thick saprolite zone. Gibbsite occurs as a major constituent locally in the B and C soil horizons in the saprolite on the plateaus, but it does not occur as a major constituent in soils on slopes, valley, or plateau areas where there has

Table 1. Description of mineralogy and granodiorite saprolite locations nos. 1 and 2.

Sample No.	Depth (meters)	Mineral composition in order of abundance			Description and remarks
		Major	Minor	Trace	
1 A	0.1 to 0.4	kaolinite gibbsite	quartz illite	orthoclase plagioclase	B soil (argillaceous horizon. Clay fraction.
1 B	0.5 to 2.0	gibbsite kaolinite	illite quartz	orthoclase plagioclase	C 1 soil horizon, soft saprolite. Clay fraction.
1 C	0.5 to 1.5	gibbsite orthoclase illite	quartz	kaolinite plagioclase	C 2 soil horizon, hard saprolite.
1 D	3.1	plagioclase quartz	orthoclase biotite		Unweathered granodiorite.
2 A	1.0 to 4.0	kaolinite	illite quartz		C 1 soil horizon, soft saprolite. Clay fraction.
2 B	3.3	plagioclase quartz	orthoclase illite	kaolinite	Very slightly weathered granodiorite. Collected 50 meters north of sample 2 A.

been extensive erosion.

Two typical exposures of granitic-gneiss and biotite-gneiss saprolite exposed on the Ashland Plateau east of Rockford in Coosa County were sampled in detail. Adams and others (1926) show the outcrops as Pinckneyville Granite, but Gault (1945) shows that the granitic rocks are mainly quartz diorite rather than granite. Orthoclase occurs as a minor constituent in the granitic phase indicating that the exposures may be classified as granodiorite gneiss rather than quartz diorite. These profiles are listed as nos. 3 and 4, and are located on Figure 1. The surface at sample profile no. 3 is about 245 meters above msl, and the surface at sample profile no. 4 is 225 meters above msl. Both locations are low rises on the plateau surface. The descriptions and clay mineralogy are given in Table 2. Both outcrops were subjected to the same weathering as profile 1 (Figure 3).

Hard sill-like masses of partly weathered rock crop out in the soft, micaceous saprolite. The hard ribs shown in the photograph were granodiorite gneiss prior to weathering, but the plagioclase has weathered to gibbsite. The adjacent areas are more gneissic and the original rock contained more biotite. Relict textures in the saprolite indicate that feldspar was a major constituent. Granodiorite gneiss grades into biotite gneiss that contains plagioclase as a minor constituent.

The gibbsite content of the saprolite is proportional to the plagioclase content of the original rock. The highest gibbsite content occurs in hard, granodiorite saprolite, and the lowest gibbsite content occurs in biotite-gneiss saprolite. Kaolinite was detected in all samples, but it is the main clay (size) constituent in the micaceous saprolite derived from the biotite gneiss. Small white clay seams, 1 to 2 centimeters

Table 2. Description and mineralogy of granite-gneiss complex saprolite, locations 2 and 4.

Sample No.	Depth (meters)	Mineral composition in order of abundance			Description and remarks
		Major	Minor	Trace	
3 A	0.2 to 1.0	gibbsite kaolinite mixed clay	quartz illite		B soil (argillaceous) horizon. Clay fraction over feldspathic granite gneiss.
3 B	1.0 to 2.8	gibbsite	mixed clay kaolinite	illite quartz	C 1 soil horizon, soft, feldspathic granite gneiss, saprolite. Clay fraction.
3 C	1.5 to 2.1	gibbsite mixed clay quartz	illite plagioclase	kaolinite	C 2 soil horizon, hard feldspathic granite gneiss saprolite.
3 D	5.0	plagioclase quartz	orthoclase gibbsite		Feldspathic gneiss, slightly weathered.
3 E	1.0 to 2.0	kaolinite	gibbsite mixed clay illite	quartz	C 1 soil horizon, soft biotite gneiss saprolite. Clay fraction.
3 F	1.0 to 2.0	gibbsite illite	plagioclase	kaolinite	White seam in granite gneiss saprolite.
3 G	1.0 to 1.5	orthoclase	muscovite	kaolinite	Small pegmatite vein in feldspathic gneiss. Adjacent saprolite contains abundant gibbsite.
4 A	0.3 to 0.9	gibbsite mixed clay kaolinite	illite orthoclase quartz	plagioclase	B soil (argillaceous) horizon, clay fraction over feldspathic granite gneiss. Clay fraction.
4 B	1.0 to 2.7	gibbsite mixed clay kaolinite	illite orthoclase quartz		C 1 soil horizon, soft feldspathic granite gneiss saprolite. Clay fraction.
4 C	1.5 to 2.5	gibbsite quartz	orthoclase illite	kaolinite	C 2 soil horizon, hard feldspathic granite gneiss saprolite.
4 D	1.0 to 2.5	kaolinite	illite mixed clay quartz	gibbsite (?)	C 1 soil horizon, soft biotite gneiss saprolite. Clay fraction.
4 E	1.0 to 2.5	kaolinite	mixed clay quartz	illite gibbsite	C 1 soil horizon, soft biotite gneiss saprolite. Clay fraction.

wide, consisting of gibbsite and clay mica, occur in the soft granitic saprolite.

Interlayered illite-vermiculite and clay-mica (illite) occur in samples derived from biotite gneiss and from granitic rocks that contained a substantial amount of biotite. Field relationships indicate that both the interlayered illite-vermiculite and the clay-mica are intermediate weathering stages between biotite and kaolinite.

The core of a small pegmatite dike that penetrated the gibbsite horizon of granitic gneiss saprolite at location no. 3 is composed of massive orthoclase bordered by coarse muscovite cyrstals. The pegmatite was slightly weathered with kaolinite partly replacing orthoclase. No gibbsite was detected, although gibbsite is the most abundant clay (size) mineral in the adjacent rocks. Since both the pegmatite and granodiorite gneiss have been subjected to the same weathering, the difference in weathered products is attributed to the difference in texture and mineralogy of the original parent rock.

A series of samples were collected from weathered rock in a small valley below the old Ashland Plateau to determine the extent of weathering below the Ashland surface (location no. 5). Sample profile

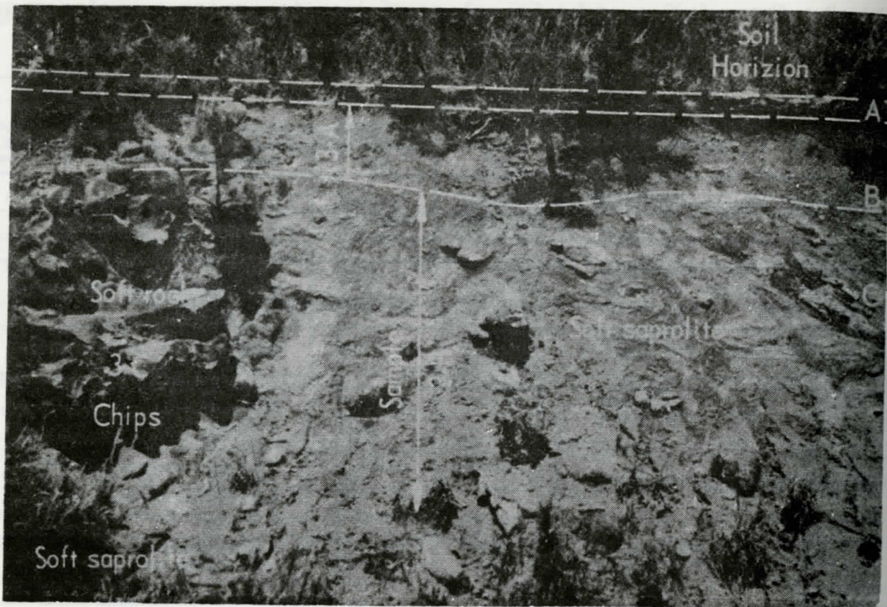

Figure 3. Granite-gneiss saprolite, location no. 3.

no. 5 was collected from a deeply weathered, well-drained exposure on a gentle slope at the edge of the valley. The altitude, about 205 meters above msl, is about 25 meters below the plateau surface in this region. Gibbsite occurs as a minor constituent in partly weathered feldspathic granite gneiss, and it was detected as a trace constituent in biotite-gneiss saprolite. Kaolinite, interlayered illite-vermiculite, and clay mica (illite) are major constituents of the granodiorite saprolite, but kaolinite is the only major clay size constituent in the biotite-gneiss saprolite.

CONCLUSIONS AND COMPARISONS WITH OTHER DISTRICTS

Topographic Controls

The results of this study show that gibbsite occurs in the saprolite under the old plateau surfaces. Gibbsite is controlled by drainage because the plateaus are better drained than similar soils found in the valleys. Because gibbsite and quartz are not mutually stable in a closed system, where there is poor or restricted drainage, gibbsite can remain associated with quartz only if solutions that are dissolving silica are drained out of the system (Garrels and Christ, 1965; Kittrick, 1969). Otherwise gibbsite is resilicated to kaolinite.

The high plateau saprolites may be a weathering product of a past climate that was warmer and more humid than the present climate. Lamb (1968) describes Pliocene age vertebarate fossils near Mobile, Alabama, that indicate a tropical or subtropical humid climate. It is difficult to determine if gibbsite is being formed in Alabama rocks now, but rocks that were weathered during the Pleistocene, or later, do not contain significant amounts of gibbsite.

Source Rocks and Weathering Paragenesis

The gibbsite content of saprolite samples collected in Alabama is directly related to the plagioclase content of the parent rock. This is in agreement with early work by Harrison (1931) in Guiana who found that plagioclase feldspar weathers directly to gibbsite in a humid tropical environment, but the gibbsite resilicates to kaolinite where drainage is restricted. Orthoclase feldspar alters to kaolinite under the same environmental conditions that the plagioclase feldspar weathers to gibbsite. This also confirms Harrison's conclusions based on studies in Guiana and Mead (1915) in Arkansas who found that with continued laterization, kaolinite is further desilicated to gibbsite. Clay-mica and interlayered illite-vermiculite replace biotite in partly weathered biotite-gneiss, but kaolinite occurs in place of biotite in the more intensely weathered biotite gneiss. Field relationships indicate that clay-mica and interlayered illite-vermiculite are intermediate weathering stages between biotite and kaolinite.

Both plagioclase and orthoclase feldspars alter to kaolinite where there is restricted drainage. Gibbsite was not detected in poorly drained areas such as location no. 2 where kaolinite was found replacing plagioclase. It was not determined during this survey if the plagioclase feldspar goes through a gibbsite stage during the destruction of the feldspar lattice, but the ultimate effect is that where there is abundant silica in solution, both feldspars weather to kaolinite.

The correlation of gibbsite in saprolite with the plagioclase content of the original rock explains the erratic occurrence of gibbsite found overlying biotite gneiss, granitic gneiss and migmatites in the Greenville-Spartanburg, South Carolina, gibbsite prospect areas. A sample that includes an altered plagioclase stringer or porphyroblast could contain more than 30 percent gibbsite, whereas a sample collected from biotite-gneiss saprolite less than a meter away might contain only a trace (less than 5 percent) of gibbsite.

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HISTOLOGY AND PETROLOGY OF TRIASSIC FOSSIL VERTEBRATE BONE

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ABSTRACT

Triassic vertebrate bone remains from the lower Comnock Formation of the Deep River Triassic Basin of North Carolina were examined by x-ray diffraction, chromatography and thin section techniques, to determine their method of fossilization.

Histologic features at the microscopic level were retained from the pre-fossilized bone. Organic matter was present but not critical to fossilized structure, while crystalline apatite, induced from outside, was the fossilizing and structure retaining media.

INTRODUCTION

Fossil vertebrae and other well preserved bone material has been found in a coal prospect area in the Triassic Deep River Coal Field located at the junction of North Carolina Road 2140 and U. S. 421, approximately 12 miles northwest of Sanford, North Carolina. The remains are associated with rocks overlying the Comnock-Gulf Coal measures (Reinemund, 1955) and are located stratigraphically about two hundred feet above the Pekin-Comnock Formation boundary. The fossil bones would, therefore, be lower Comnock and late Triassic in age.

The external morphology and internal structures are well preserved in many of the specimens, even at the microscopic level. Primary and secondary bone relationships remain intact as do other histologic features that are unique to the specific bone type.

The excellent preservation of such features from Triassic time lends itself to a study of the preservation processes involved. The primary concern of this paper is, therefore, to investigate the reasons and methods for such preservation. Considerations of possible replacement of both organic and inorganic material, the breakdown of the organic fraction and the preservation of any organic residue are included.

Acknowledgments

The author is indebted to Don Miller for performing the chromatography and to Jim Boling for assistance with the microphotographs.

METHODS

Thin section studies were undertaken on untreated fossil bone and on fossil bone heated to 450°C for 12 hours. With the untreated bone, normal petrographic techniques for making thin sections were followed except that, because of intense stain of color, some sections had to be cut thinner than the normal 30 microns to permit sufficient light transmission. Heating the sample to 450°C for 12 hours reduced the organic material and removed much of the stain so that the normal 30 micron section could be used. The secondary mineralization and what remained of the original primary osseous calcification gave the sample enough rigidity, however, so that normal non-impregnating petrographic thin section cutting techniques could be used. The thin sections for both untreated and heated samples were cut transversely from small long bones. Each thin section included a cross section of both compact (including cortical compaction, primary and secondary) and cancellous bone.

X-ray diffraction patterns were made from the powder of whole fossil bone samples. Any traces not attributable to apatite were extracted and the implications of the extracted traces considered.

Chromatographic analysis were made on whole bone samples. Whole bone powder was first treated with 2 percent HCl to remove highly soluable material. It was then hydrolyzed with 6 N HCl for 36 hours or more in return condenser flasks over a boiling water bath. The HCl was distilled in vacuum with the predominance of salts precipitated by ethyl alcohol and the remainder by passing solution through a column of Dowex-50 resin in H+ form. The residue was dissolved in 10 percent isopropyl alcohol. Paper chromatography was used to determine the presence or absence of amino acids with a 4/1/5 ratio of n-butyl alcohol/ glacial acetic acid/water as the solvent.

OBSERVATIONS

Megascopic

External and internal structural forms and features are retained in those bones of the deposit that had not been subjected to severe movement or post-fossilization ground-water solution. Visual identification of the material filling the cavities is not possible at the megascopic level. Irregular-shaped post-fossilization solution openings and fractures can sometimes be seen either partially or wholly filled with an unctuous tan to pink clay-like material.

By heating sections at 450°C for 12 hours, the black osseous organic tissue was reduced to a light colored ash. This process reversed the color contrast with the original osseous tissue becoming the lighter color with limonite-stained secondary cavity fillings.

Microscopic

Normal fossil bone, under ordinary light transmission through a petrographic microscope, showed all of the gross features listed under megascopic and also revealed individual haversian systems. The osteons are outlined by their junction with primary or interstitial bone and in many cases this junction carries a prominent cement line (Figure 1). Lamellae of both primary and secondary bone can be seen (Figure 1). Lacunae are evident but usually not well defined (Figure 2).

Many of the haversian canals and the inter-trabeculae areas are partially or wholly filled with transparent secondary crystals radiating inward from the walls of the openings (Figure 2 and 3). Some contain iron stain at the junction of the crystals with the walls (Figure2). Clay occupies some of the larger inter-trabeculae openings not completely filled with transparent crystals and occurs in solution channels and solution cavities that are generally confined to the more open cancellated structure.

Under plane-polarized light transmission, the birefringent nature of the normal fossil bone material becomes evident (Figure 3). Many of the haversian systems are marked by faint pseudeo-spherite crosses (Figure 3) resulting from concentric bands or lamellae of the osteon. Where the osteon is not circular, the cross is irregular and less intense and the shape of the cross then varies with rotation of the microscope stage.

Under crossed nicols the osteons are well defined by alternating light and dark lamellae (Figure 3). The primary bone carries the same banding but in a parallel rather than a circular pattern (Figure 1). The junction of primary and secondary bone is quite evident and is accentuated by color intensity and variation in its angular relationship (Figure 1).

Secondary mineralization becomes more pronounced under polarized light because of birefringence. The pattern of apatite crystals radiating inward from the haversian canal walls of inter-trabeculae spaces as seen under ordinary light is retained, but siderite crystals can now be seen at the junction. The relationship is not clear, however, because of intense iron stain at this point.

A pattern of birefringent single crystallites that is roughly parallel to the osteon or primary bone lamellae emerges as the microscope stage is rotated (Figure 2). Because of the stain present in the osseous tissue and because of a similarity of the crystallites refractive index with that of their matrix they are neither a pronounced or consistant feature. They roughly follow the disperal pattern expected of the lacunae.

Residual osteons are present in the cortical region of the compact bone (Figure 1). An anomaly exists, however, inasmuch as in some instances the primary bone abutts the residual osteon structure while in others the lamellae of the primary osseous tissue continue around and are offset by the residual osteon (Figure 1).

When normal fossil bone was heated at 450°C for 12 hours, the previously nearly opaque organic content was changed to an unctuous white to yellow ash. The thin section of bone became transparent to ordinary light and lower in both refractive index and density. Thin sections could be cut from the heated bone material because it was held together by the interosseous tissue crystalline content.

Under ordinary light lamellae of the osteons and primary osseous tissue were faint and their presence intermittent and scattered. Because of increased transparency and reduced refractive index of the matrix material induced by heating, however, all apatite and other crystals stand out in relief and can be seen quite readily (Figure 4).

As a result, it can be seen that: 1) Many of the crystals adjacent to the haversian canal wall penetrate the wall by replacement of osseous tissue; 2) limonite or other iron oxide material is intermixed with apatite crystals adjacent to the haversian canal wall; 3) some siderite crystals are present in the same area as the iron oxides; 4) both the areas of previous primary and secondary osseous tissue contain apatite crystals. These are either euhedral hexagonal, or irregular in shape with the euhedral shaped crystals being the larger. The irregular-shaped crystals have long axis parallel to lamanae and assume the spatial relations of lacunae; 5) no canaliculi can be detected.

Under crossed nicols all crystals become visible as the stage is rotated to the proper position. Lamellae are present, though vague: and are outlined by crystals in the lacunae.

X-ray Analysis

X-ray patterns were obtained for whole bone samples and, therefore, show total composition for both the possible remaining primary and secondary crystalline products. No doubling or broadening of peaks was noted for the apatite content and it was assumed that all apatite involved, whether primary or secondary, was of the same composition. The x-ray pattern of the whole bone was similar to that which Brophy et al., (1968) found for primary bone structure. Predominance of well crystallized secondary apatite in the haversion canals and inter-trabaculae areas results in a well defined x-ray pattern with sharp, intense peaks.

Kaolinite, siderite and goethite peaks were also noted in some patterns.

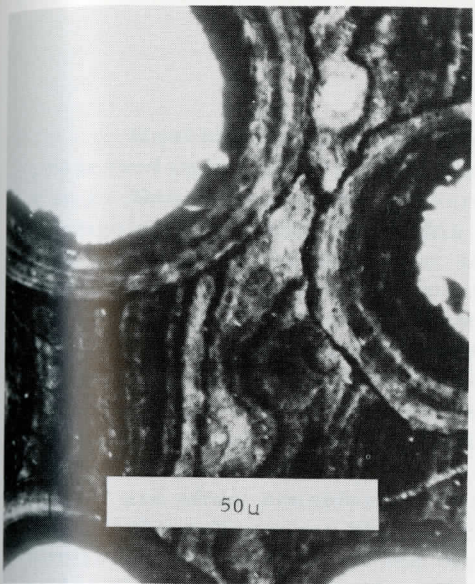

Figure 1. Thin section of compact bone that contains both primary and secondary bone areas, residual osteons, cement lines and lamella.

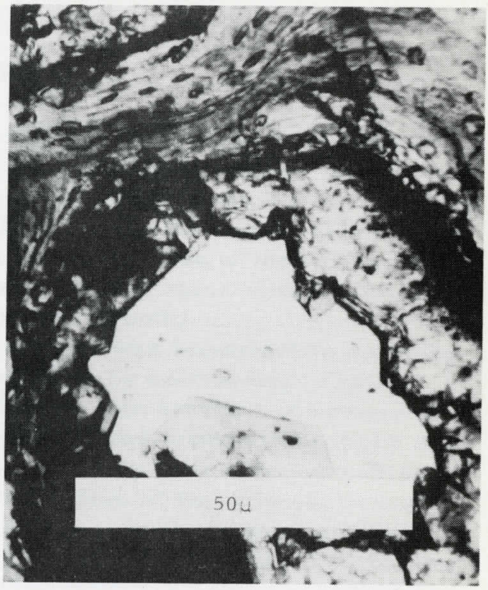

Figure 2. Thin section of an area of secondary bone from near cancelleous zone. Canal is partly filled with apatite and some siderite with its oxidation product. Lacunae fillings and large secondary isolated crystals are present.

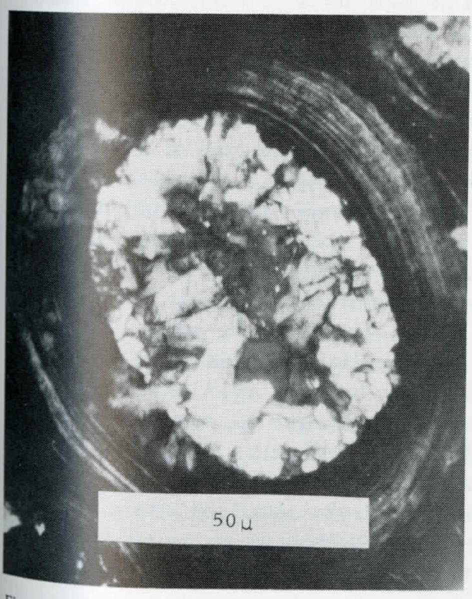

Figure 3. Thin section of haversian canal filled with secondary crystals. Lamella are preserved and lacunae indicated. A pseudo-spherite cross is present.

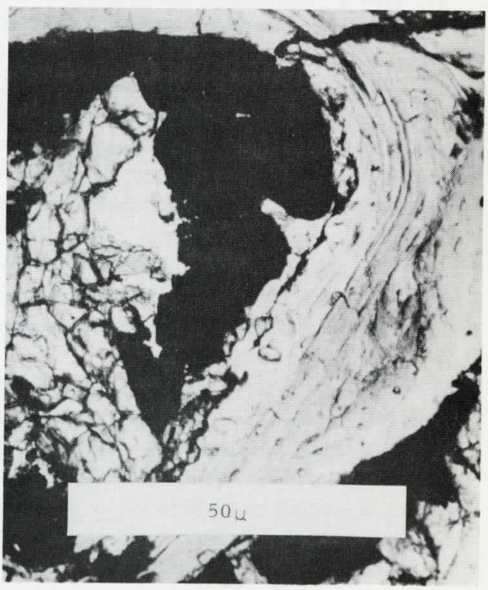

Figure 4. This section of cancelleous bone heated to 450°C/12 hours. Lamella remain visible, lacunae fillings are prominent, secondary iron oxide is present and organic material removed.

Chromatography

Negative results were obtained for the presence of any amino acids and the organic content must, therefore be of a complex nature.

DISCUSSION

Most histologic features present in fresh bone have been well preserved in the Triassic age fossil bones which are under consideration in this study. The preservation of these features is revealed through optical continuity in thin section and is, therefore; 1) a reflection of either the preservation of at least a portion of the original osseous tissue or, 2) a replacement that gives the same optice effects as the original structure or, 3) a combination of both.

That the presence of organic residue was important in outlining these histologic features and can be verified by contrasting Figures 1 or 3 with Figure 4. Figures 1 and 3 show thin sections of normal untreated fossil bone specimens. Alternating and contrasting light and dark bands outline both primary and secondary bone structure (Figure 1) and are often considered to be a reflection of the original arrangement of the collogenous fibers. Figure 4 represents a thin section of a similar specimen in which the organic portion was destroyed by heating to 450°C/12 hours. Virtually all coloring matter has been removed except for fracture filling and secondary crystals of inorganic oxide compounds.

Megascopically the osseous tissue of the specimens appears dark and opaque. In a thin section of normal 30 micron thickness it also appears black and nearly opaque. As the thickness of the section is reduced, however, it becomes a transparent brown to red-brown and with magnification resolves into light and dark bands (Figure 1 and 3). Cook et al., (1962) considered similar coloring, found in their study, to be the result of soil solute penetration and subsequent stain. They noted a stain intensity gradation from dark to light from the exterior toward the interior of their specimens and considered this proof of foreign matter penetration. No such color intensity gradation occurs in the specimens of this study. Furthermore, if variation in porosity between band units did exist so that differential penetration of soil solute could take place the consistancy of the banding precludes such a mechanism (Figure 1) as an explanation.

In the "age series" of fossil bones studied by Cook et al., (1962) they found that all organic bone tissue was deplete by Jurassic and Triassic time with complete mineralization of the specimens having taken place. This conclusion carried an implication of time which is refuted by the more realistic views of Drozdova et al., (1966). They suggest that amino acids resulting from breakdown of the organic portion of the osseous tissue tends to be lost or destroyed during cementation or chemical redistribution of materials during diagenesis and that the process is dependent on neither degree of fossilization, as pointed out by Ableson (1963) nor on age. They found a varied amino acid content in fossil bone back to upper Permian time and concluded that preservation of organic material and amino acids is a function of burial in a host material where no redistribution of material occurs. They considered uncemented sands the best host to fulfill this requisite. The bones of this study were found in clay or argillite of high organic content. Secondary mineralization from solution was prevalant as attested to by crystal deposition in the canals and inter-trabacular openings. Ground water solution and filling of fractures and solution voids were present but occurred later than fossilization and, therefore, were not involved in the fossilization process.

Organic material was definitely present in a degraded state but not in amounts to lend structural rigidity to the bone remains. An extension or bridging of the original crystalline framework was necessary to retain the rigidity. Optical continuity of the original histologic organization was retained, therefore, the extension must have taken place in an orderly fashion contiguous with the original structure. This occurred as a replacement of the organic matter or, more likely, as an extension of original structure on pre-existing mineral matter as deterioration of organic material took place. Such a process is indicated by the retention of the original optic pattern furnished by the pre-existing crystalline matter.

Three other distinct modes of secondary crystallization were evident within the framework of the fossil bone structure.

Most prominent of these was that of partial or complete filling of canals of the haversian systems (Figure 3) or inter-trabecular voids of the spongy bone (Figure 2). Apatite is the predominant mineral material of such fillings and is deposited from the peripheral surfaces inward toward the center of the opening. In some instances the inner lamella of the osteons appear to be replaced by the first deposited crystals (Figure 3). In some instances hematite and siderite are deposited with the apatite (Figure 2). The deposition of siderite appears to be simultaneous with apatite. The hematite, however, appears to be a later oxidation product of the siderite with cross cutting relationships extending into late fractures of the bone.

The second mode takes the form of either crystals or crystal complexes with their long axis extending parallel to the lamellar striations of either primary or secondary osseous tissue (Figure 2 and 4). Their position and number (where developed) are similar to that of the lacunae filling; either partially, wholly or in some cases over extended.

The third mode results in the development of large isolated euhedral hexagonal shaped crystals. They are randomly spaced but conform to the lamellar pattern of the bone structure (Figure 2 and 4). They could be either over extended lacunae fillings, possibly viewed from their flat orientation, or crystals that grew in-situ and which are not dependent upon lacunae for their inception or development. If the latter is the case, a question would arise as to why they are larger or why they even exist as they are not associated with visible fracture patterns or transport channels. Cook et al., (1962) noted such crystals but considered them to be pre-existing crystals or crystal fragments that became embedded within the bone structure during growth. They gave no source or reason for the particles presence before inclusion into the bone structure.

Cook et al., (1962) also found canaliculai in Jurassic brontosaurus bone well defined by brown pigment content. Canaliculi could not be recognized in the present study either as voids or as inorganic or organic filling. It is suspected that they are filled with apatite oriented to the surrounding material and, therefore, cannot be delineated with the petrographic microscope.

Because the lacunae are centers of secondary crystallization, it would seem plausible that the associated canaliculi acted as passageways, not only for the filling of the lacunae but for transportation of material in the extended crystallization of the bone matrix. This extension of the original inorganic units of the osseous tissue has preserved the histologic pattern in minute detail. Rogers and Nash (1924) pointed out, that there is little evidence of the migration of calcium phosphate away from fossil bones. Considering this together with the facts of this study, it is concluded that, in this case the fossilization process is one of preservation and extension of the original inorganic content which preserves the original histologic pattern.

CONCLUSIONS

Organic matter of the initial collogenous fibers is partially retained but is not preserved in its original form. Its breakdown product is represented by a heavy stain.

The calcification pattern of the original osseous tissue is preserved in minute detail. This is accomplished through extension of the original crystalline units and reflects a controlled continuity with the original calcification pattern. The chemical composition of the original apatite and the extensional material appear to be the same.

Lacunae and canaliculai appear to be the basic transport system involved in extension of the original osseous tissue inorganic units in the process of bone fossilization.

Both extension of the osseous tissue crystalline units and filling of haversian canals and inter-trabeculae opeing, with secondary mineral material indicates a ready source for calcium and phosphorous bearing solution. The presence of siderite within the depositional sequence would point to the existence of a depositional condition within a limited range of eh-pH concomintant with siderite stability (Krauskopf, 1967).

Post-fossilization ground water activity created solution cavities, made oxygen available, and infilled with secondary mineral and detrital material.

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THORIUM, URANIUM, AND POTASSIUM IN UPPER CRETACEOUS PALEOCENE AND EOCENE SEDIMENTS OF THE LITTLE TALLAHATCHIE RIVER WATERSHED IN NORTHERN MISSISSIPPI[1/]

By

Jerry C. Ritchie, Paul H. Hawks, J. Roger McHenry[2/]

ABSTRACT

Gamma-ray spectrometric analyses were made of 129 samples collected from outcrops of ten geologic formations in north Mississippi. Thorium, uranium and potassium-40 concentrations were determined. Thorium concentrations ranged from 0.9 to 29.9 ppm, uranium from 0.3 to 10.9 ppm and potassium from nondetectable to 2.43 percent. Low concentrations of these elements were found in the sands of Kosciusko and Meridian Formations (Eocene) and the McNairy Sand Member of the Ripley Formation (Upper Cretaceous). Concentrations were three to five times greater in the fine-textured formations.

INTRODUCTION

Thorium, uranium, and potassium-40 are naturally occurring radionuclides that have been used in stratigraphy for the purpose of age determination and correlation. Well logs of gamma-ray activity (Broda and Schonfeld, 1966; Barsukov <u>et. al.</u>, 1965) and aerial gamma-ray surveys (Schmidt, 1962; MacKallor, 1963) correlate with changes in geologic materials and have been used for geologic mapping. In a given mapable lithologic unit, the distribution of radionuclides may be fairly uniform and of sufficiently diagnostic character and quantity to be a useful tool in geologic investigations. Frequently, terrestial deposits are

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2/ Botanist, Geologist, Research Chemist, respectively, Oxford, Mississippi.

complex and the geologist encounters lateral and vertical changes. In order that radionuclide data be meaningful, detailed information of the lithologic characteristics and relationships may be required.

The purpose of this study was to determine thorium, uranium and potassium concentrations of different known sedimentary formations and to correlate these geochemical data with other data on the characteristics of the formations.

METHODS AND MATERIALS

The study area was the Little Tallahatchie River watershed upstream from Sardis Dam in northern Mississippi (Figure 1). Located in the Coastal Plain Province, the geologic material is composed of Upper Cretaceous, Paleocene and Eocene sediments. Loess covers the watershed in depths ranging from 0 to 3 meters. The geologic formations dip to the west. The formations from east to west are Selma, Ripley (McNairy Member), Prairie Bluff, Clayton, Porters Creek, Fearn Springs, Ackerman, Meridian, Tallahatta and Kosciusko (Attaya, 1951; Conant, 1941, 1942; Lusk, 1956; Priddy, 1943, Vestal, 1954, 1956).

Samples were collected from exposed road cuts or eroding gullies. Sites were selected to give maximum exposure of the geologic column. The samples were brought into the laboratory, sieved through a 12-mm screen to insure uniform aggregate size and dried. The dried sample (2500 grams) was put in Marinelli counting beakers. The beakers made counting geometry and sample density constant for all samples. The beakers were sealed and stored for a minimum of two weeks so that thorium and uranium would reach equilibrium with their daughter products before gamma-ray spectrometric analysis was made.

Gamma-ray spectrometric analyses were made using a 1024-channel pulse height analyzer in conjunction with a 10- x 12.5-cm thallium-activated sodium iodide crystal (Ritchie et. al., 1970). The gamma-ray spectra were reduced by an IBM-360-40 computer using a least squares technique for fitting spectra (curves) of standards of unknown sample spectra to determine concentration (Schonfeld, 1966). Thorium and uranium concentrations are expressed in parts per million (ppm) and total potassium is expressed in percent (%) by dry weight. Confidence limits at the .05 level of probability for thorium determinations was $\pm$ 4%, for uranium $\pm$ 6%, and for potassium $\pm$ 3%.

Mechanical analyses were made using a modification of the pipette method of Kilmer and Alexander (1949).

RESULTS AND DISCUSSION

The measured concentrations of thorium, uranium and potassium

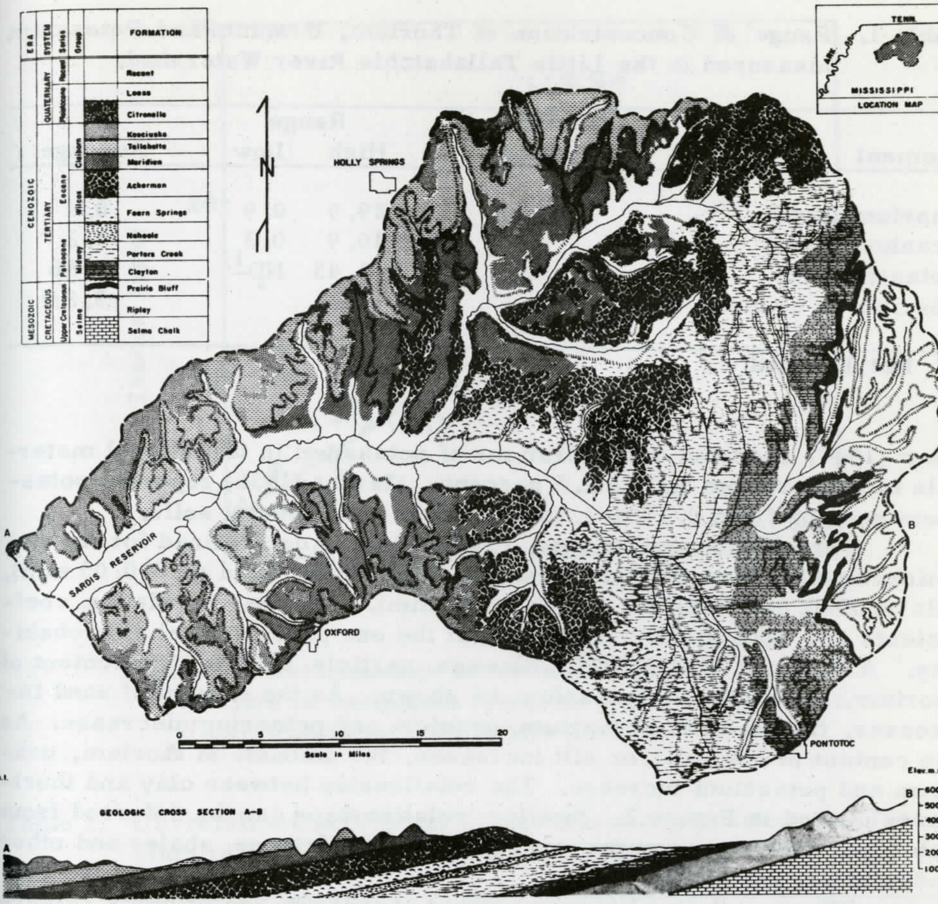

Figure 1. Geologic map of the Little Tallahatchie River Watershed upstream from Sardis Dam, northern Mississippi.

and particle size distribution are given in Appendix I. Thorium concentrations ranged from 0.9 to 29.9 ppm with an average of 8.4 ppm (Table 1). Uranium concentration ranged from 0.3 to 10.9 ppm with an average of 2.2 ppm. The average thorium to uranium ratio of 3.8:1 is within the theoretical ratio of 3:1 to 4:1 proposed by Adams et. al. (1959) for accessible terrestrial materials and the same as the 3.8:1 ratio proposed by Mahdavi (1964) for a "homogenized, well-mixed and representative sample of all crustal environments of mineral formation". The concentrations of thorium and uranium are within the ranges reported by Ritchie and Plummer (1969) for similar material from the Georgia Coastal Plain.

Potassium concentrations ranged from nondetectable levels to 2.43 percent, with an average of 0.86 percent. Clark, Anderson, and

Table 1. Range of Concentration of Thorium, Uranium and Potassium Measured in the Little Tallahatchie River Watershed.

Element	Number of Samples	Range High	Range Low	Average
Thorium (ppm)	129	29.9	0.9	8.4
Uranium (ppm)	129	10.9	0.3	2.2
Potassium (%)	129	2.43	ND1/	0.86
Thorium to uranium ratio				3.8:1

1/ Not detected

Shaw (1944) reported concentrations of potassium in terrestrial materials to ranges from 0.3 to 2.5 percent. Marbut (1935) reported potassium ranging from 0.17 to 3.30 percent in agricultural soils.

Correlation coefficients, (Table 2) were determined for the relationships between thorium, uranium, potassium, sand (2.0-0.05 mm), silt (0.05-0.002 mm) and clay (< 0.002 mm). All the correlation coefficients are statistically significant at the one percent level of probability. A definite relationship between particle size and the content of thorium, uranium and potassium is shown. As the content of sand increases, the amounts of thorium, uranium and potassium decrease. As the content of clay and/or silt increases, the amounts of thorium, uranium and potassium increase. The relationship between clay and thorium is plotted in Figure 2. Similar relationships can be inferred from published reports on gamma-ray activity of sandstones, shales and other sedimentary materials.

The correlation between content of thorium and uranium (Figure 3) is positive, indicating that thorium and uranium probably have similar geochemical cycles in these sedimentary materials as opposed to the general geochemistry of magmatic deposits (Olson and Overstreet, 1964). Adams, Osmond and Rogers (1959) have reported on the general similarity of the geochemical cycles of thorium and uranium. No significant relationships were found between thorium, uranium and potassium concentrations and nitrogen, organic matter and carbon content and pH. Potassium concentrations were independent of thorium and uranium concentrations, indicating the geochemical cycle for potassium in these samples was different from that for thorium and uranium.

The concentrations of thorium, uranium and potassium differed in the various geologic formations (Table 3). Gamma-ray activity was less in sandy formations than in finer-textured formations. The Kosciusko, Meridian and McNairy Members of the Ripley formation are sandy. These three sandy formations have thorium, uranium and potassium concentrations comparable to the Gulf Coast sands studied by Mahdavi (1964), and had gamma-ray activities distinguishable from those of

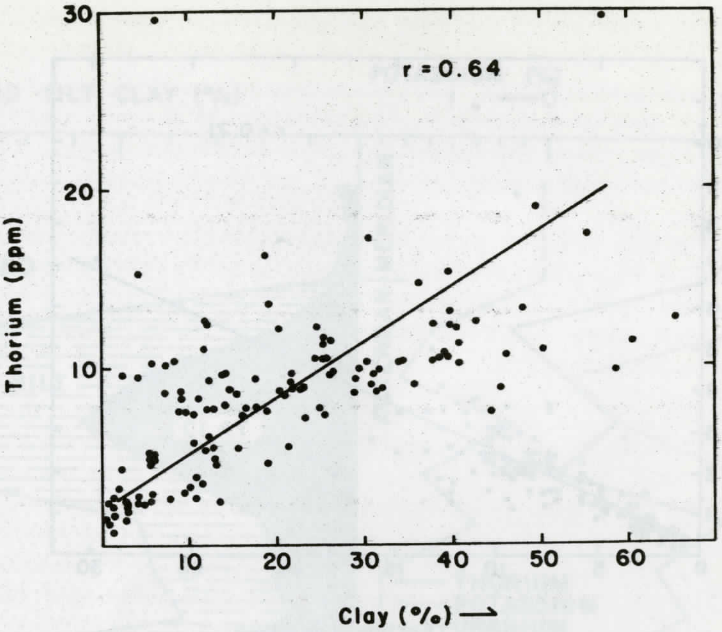

Figure 2. Relationship between contents of clay and thorium in sediments from the Little Tallahatchie River Watershed, north Mississippi.

Table 2. Correlation Coefficients (r) for Thorium, Uranium and Potassium in Relation to Sand, Silt and Clay.

	Thorium	Uranium	Potassium
Sand	-0.64	-0.59	-0.70
Silt	0.51	0.49	0.71
Clay	0.64	0.57	0.52
Thorium	---	0.71	0.44
Uranium	---	---	0.36

the other formations with which they are in contact.

The finer textured formations (Tallahatta, Ackerman, Fearn Springs, Porters Creek, Clayton, Prairie Bluff, and Selma) have thorium and uranium concentrations approximately three times those found in the sandy formations and potassium concentrations approximately 5-6 times those of the sandy formations. These finer textured formations can be distinguished from the sandy formations by their gamma-ray activity. However, the gamma-ray activity of these finer-textured

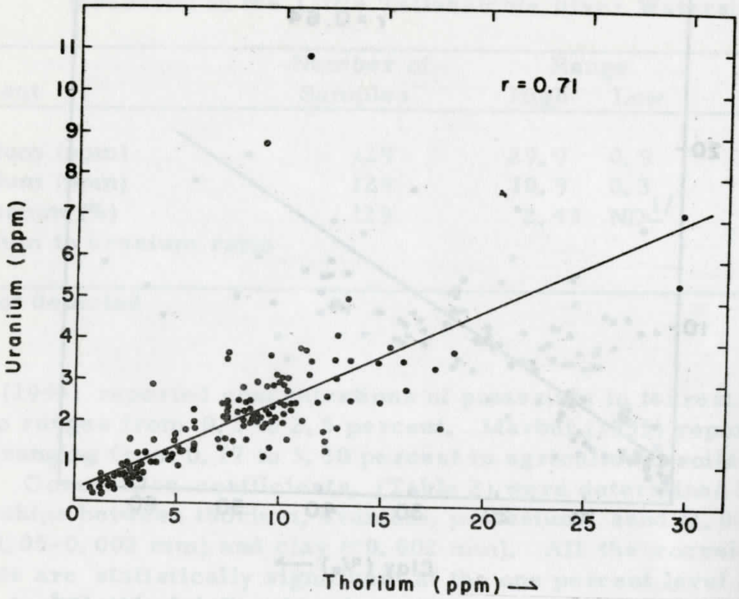

Figure 3. Relationship between contents of thorium and uranium, in sediments from Little Tallahatchie River Watershed, north Mississippi.

Table 3. Thorium, Uranium and Potassium Concentrations in the Geologic Formation of the Little Tallahatchie River Watershed.

Geologic Formation	Thorium ppm	Uranium ppm	Th/U1/ Ratio	Potassium %
Kosciusko	3.09	0.86	3.9:1	0.12
Tallahatta	11.10	2.52	4.4:1	0.75
Meridian	2.76	0.73	4.0:1	0.12
Ackerman	10.08	2.80	4.1:1	1.17
Fearn Springs	8.77	2.43	3.7:1	1.47
Porters Creek	10.81	2.35	5.4:1	1.77
Clayton	9.01	2.04	4.7:1	0.69
Prairie Bluff	9.19	2.28	4.0:1	1.00
Ripley (McNairy member)	3.68	1.14	3.2:1	0.29
Selma	7.77	2.40	3.3:1	1.32

1/ Average of individual ratios

formations is so similar that it is difficult to distinguish differences between any two fine-textured formations.

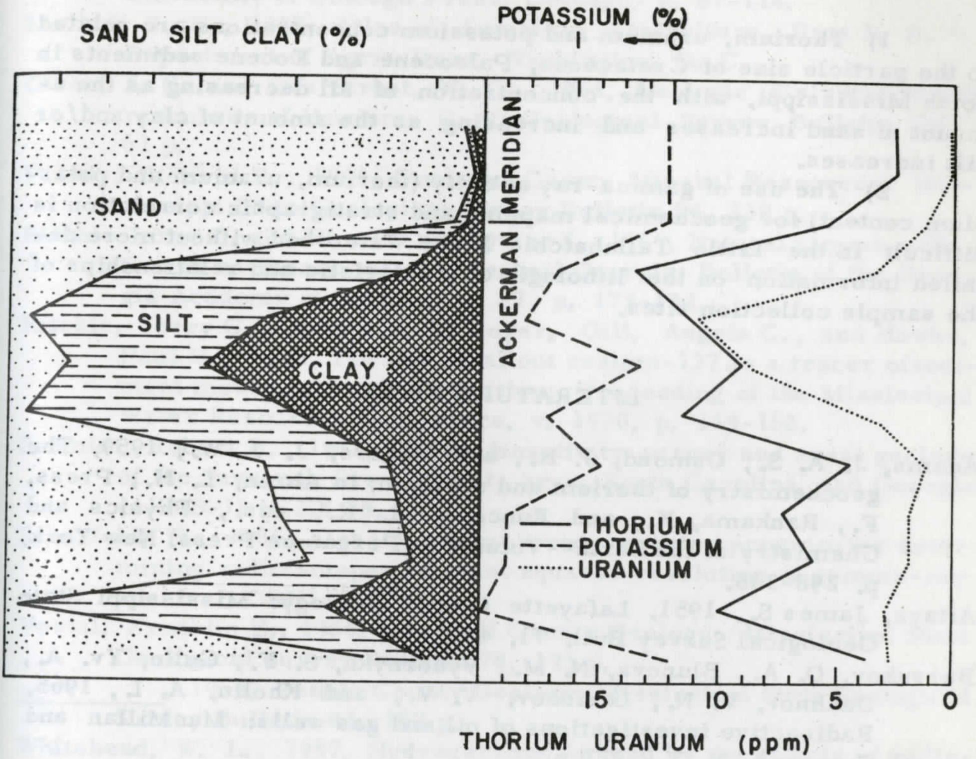

Figure 4. Particle size, thorium, uranium and potassium relationship in a 120-foot vertical section of the Meridian and Ackerman formations.

Concentrations of uranium and potassium measured in the Selma formation are lower than the 7.9 ppm uranium and 1.91 percent potassium reported for a Selma formation sample from Mississippi (Whitehead, 1957). The highest concentrations of thorium and uranium were measured in two undifferentiated Wilcox samples. Thorium concentrations of 29.2 and 29.9 ppm and uranium concentrations of 4.4 and 7.2 ppm were measured in samples 70-074 and 70-075. Grim (1936) has reported zircon and monazite to be present in the Wilcox Group of the Little Tallahatchie River Watershed area which could account for higher thorium and uranium concentrations.

The variation in content of thorium, uranium, potassium, sand, silt and clay in a 120-ft. vertical section of Meridian and Ackerman is shown in Figure 4. The variation in the Ackerman Formation and the sharp break between the Ackerman and Meridian Formations is shown by both the gamma-ray activity and the particle size relationship.

CONCLUSIONS

1) Thorium, uranium and potassium concentrations are related to the particle size of Cretaceous, Paleocene and Eocene sediments in north Mississippi, with the concentration of all decreasing as the amount of sand increases and increasing as the amount of clay and/or silt increases.

2) The use of gamma-ray activity (thorium, uranium and potassium content) for geochemical mapping and stratigraphic correlation is difficult in the Little Tallahatchie River Watershed without more detailed information on the lithologic characteristic and relationships of the sample collection sites.

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Appendix I: Thorium, Uranium and Potassium Concentration and Particle Size Measured in Geologic Material from the Little Tallahatchie River Watershed, Mississippi.

Sample No.	Geologic Material	Thorium (ppm)	Uranium (ppm)	Th/U Ratio	Potassium %	Sand1/ %	Silt2/ %	Clay3/ %
69-012	Wilcox4/	7.33	2.22	3.3:1	1.28	44	31	25
69-013	Wilcox4/	10.58	3.08	3.4:1	1.94	8	45	47
69-014	Wilcox4/	10.17	2.94	3.5:1	1.12	33	26	41
69-018	Porters Creek	9.47	3.74	2.5:1	1.73	23	73	4
69-033	Meridian	2.84	0.65	4.4:1	0.04	93	2	5
69-034	Ackerman	8.52	1.98	4.3:1	0.73	51	34	15
69-035	McNairy	1.85	0.62	3.0:1	0.30	98	1	1
69-037	Fearn Springs	8.38	2.45	3.4:1	1.64	37	34	29
69-038	Fearn Springs	8.86	2.91	3.0:1	1.84	57	35	8
69-039	Porters Creek	7.64	2.16	3.5:1	0.96	47	32	21
69-040	Porters Creek	12.53	1.31	9.6:1	2.01	4	30	66
69-041	Ripley5/	1.78	0.50	3.6:1	0.01	97	2	1
69-042	Ripley5/	2.43	0.85	2.9:1	0.04	91	1	7
69-043	Ackerman	12.30	2.55	4.8:1	1.11	7	52	41
69-044	Prairie Bluff	9.19	2.28	4.0:1	1.00	53	18	29
69-045	Selma	8.06	2.70	3.0:1	1.30	22	33	45
69-046	Selma	7.10	2.43	2.9:1	1.23	20	35	45
69-047	Selma	8.14	2.07	3.9:1	1.43	19	35	46
69-048	Wilcox4/	12.70	1.95	6.5:1	0.31	72	3	25
69-049	Clayton	16.11	2.80	5.6:1	1.71	65	17	18
69-050	Wilcox4/	9.22	1.24	7.4:1	0.47	71	14	15
69-051	Wilcox4/	17.94	1.86	9.6:1	0.57	40	29	31
69-052	Ripley5/	2.98	0.78	3.8:1	0.15	82	5	13
69-053	Porters Creek	9.84	3.12	3.2:1	1.76	30	65	5
69-172A	Tallahatta	10.50	2.27	4.6:1	0.82	8	53	39
69-172B	Tallahatta	10.68	2.42	4.4:1	0.87	7	54	39
69-175A	Tallahatta	10.29	2.42	4.3:1	0.70	21	45	34
69-175B	Tallahatta	10.04	2.51	4.0:1	0.64	22	44	34
69-191	Meridian	3.77	0.77	4.9:1	0.09	84	3	13
70-002	Ackerman	9.40	2.21	4.3:1	0.63	60	28	12
70-003	Ackerman	10.22	2.17	4.7:1	1.01	29	45	26
70-004	Porters Creek	9.60	1.38	7.0:1	1.47	20	52	28
70-005	Porters Creek	14.94	2.47	6.0:1	2.15	15	49	36
70-006	Fearn Spring	10.24	2.58	4.0:1	1.44	7	69	24
70-007	Fearn Spring	7.40	2.46	3.0:1	0.56	67	20	13
70-008	Ackerman	8.35	1.46	5.7:1	1.75	41	36	23
70-009	Ackerman	9.25	1.86	5.0:1	1.83	36	37	27
70-010	Ackerman	8.69	2.26	3.9:1	1.22	58	19	23
70-011	Ackerman	8.27	2.11	3.9:1	1.22	59	19	22
70-012	Ackerman-Meridian	4.83	1.32	3.7:1	0.68	75	12	13
70-013	Ackerman-Meridian	5.06	1.34	3.8:1	0.71	75	12	13
70-014	Fearn Spring	7.89	2.34	3.4:1	0.94	41	42	17
70-015	Fearn Spring	9.47	2.38	4.0:1	1.75	45	32	23
70-016	Ackerman	9.95	3.02	3.3:1	1.32	23	59	18
70-017	Ackerman	8.46	2.50	3.4:1	1.11	34	51	15
70-018	Meridian	2.71	0.53	5.1:1	0.05	93	3	4
70-019	Ackerman	9.57	2.09	4.6:1	0.43	52	27	21
70-020	Ackerman	8.87	2.07	4.3:1	1.79	42	35	23
70-021	Ackerman	8.60	2.18	3.9:1	1.84	27	41	32
70-022	Meridian	5.00	0.76	6.6:1	0.07	92	2	6
70-023	Tallahatta	12.84	3.12	4.1:1	0.82	70	18	12
70-024	Meridian	1.66	0.56	3.0:1	0.02	96	1	3
70-025	Tallahatta	18.43	3.73	4.9:1	1.11	6	44	50
70-026	Kosciusko	4.80	1.24	3.9:1	0.34	76	10	14
70-027	Meridian	2.12	0.67	3.2:1	0.06	97	2	1
70-028	Ackerman	17.53	3.35	5.2:1	1.20	2	43	55
70-029	Meridian	3.29	0.73	4.5:1	0.16	96	3	1
70-030	Meridian	1.55	0.48	3.2:1	0.03	96	2	2
70-031	Meridian	0.89	0.27	3.3:1	6/ND	99	0	1
70-032	Meridian	4.04	0.96	4.2:1	0.04	96	1	3
70-033	Ackerman	9.29	2.03	4.6:1	0.84	72	15	13
70-034	Meridian	3.14	0.71	4.4:1	0.09	88	1	11
70-035	Ackerman	11.85	2.35	5.0:1	0.70	30	43	27
70-036	Meridian	1.36	0.32	4.3:1	0.03	95	1	4

Appendix I: Thorium, Uranium and Potassium Concentration and Particle Size Measured in Geologic Material from the Little Tallahatchie River Watershed, Mississippi (Continued).

Sample No.	Geologic Material	Thorium (ppm)	Uranium (ppm)	Th/U Ratio	Potassium %	Sand1/ %	Silt2/ %	Clay3/ %
70-037	Ackerman	10.18	1.98	5.1:1	1.67	34	35	31
70-038	Ackerman	8.72	2.22	3.9:1	0.82	5	59	36
70-039	Ackerman	13.39	2.50	5.4:1	0.50	64	18	18
70-040	Fearn Springs	9.16	1.90	4.8:1	2.12	20	51	29
70-041	Ackerman-Fearn Springs	8.24	2.05	4.0:1	1.74	29	38	33
70-042	Ackerman	8.96	1.95	4.6:1	1.82	35	34	31
70-043	Ackerman	7.12	1.29	5.5:1	1.37	70	13	17
70-044	Ackerman	15.95	3.43	4.7:1	0.59	89	7	4
70-045	Ackerman	14.87	3.54	4.2:1	1.20	26	34	40
70-046	Ackerman	10.85	2.44	4.4:1	2.16	13	49	38
70-047	Ackerman	11.05	2.03	5.4:1	1.23	1	37	62
70-048	Ackerman	12.39	3.54	3.5:1	1.07	6	73	21
70-049	Ackerman	13.23	5.06	2.6:1	0.68	2	51	47
70-050	Ackerman	10.39	2.77	3.8:1	0.73	20	42	38
70-051	Ackerman	12.70	4.22	3.0:1	1.04	17	46	37
70-052	Meridian	2.95	1.17	2.5:1	0.27	86	7	7
70-053	Ackerman	10.09	2.03	5.0:1	0.16	93	1	6
70-054	Meridian	2.53	0.55	4.6:1	0.06	92	2	6
70-055	Ackerman	10.83	3.07	3.5:1	1.00	14	61	25
70-056	Meridian	3.14	0.78	4.0:1	0.11	88	2	10
70-057	Kosciusko	3.30	1.26	2.6:1	0.11	88	4	8
70-058	Kosciusko	2.41	0.55	4.4:1	0.01	97	2	1
70-059	Kosciusko	1.85	0.39	4.7:1	0.03	95	2	3
70-060	Tallahatta	5.79	1.15	5.0:1	0.18	84	4	12
70-061	Tallahatta	5.90	1.41	4.2:1	0.16	81	6	13
70-062	Tallahatta	15.43	3.68	4.2:1	1.41	16	45	39
70-063	Ripley5/	8.66	2.12	4.1:1	0.81	88	4	8
70-064	Ripley5/	4.89	1.59	3.1:1	0.65	90	4	6
70-065	Ripley5/	5.86	2.29	2.6:1	0.38	85	5	10
70-066	Ripley5/	2.14	0.75	2.9:1	0.13	96	1	3
70-067	Ripley5/	2.51	0.78	3.2:1	0.13	95	1	4
70-068	Clayton	10.76	1.57	6.9:1	0.96	50	43	7
70-069	Clayton	8.47	2.24	3.8:1	0.35	72	19	9
70-070	Clayton	4.65	1.09	4.3:1	0.21	79	14	7
70-070B	Clayton	6.98	1.47	4.7:1	0.19	73	3	24
70-071	Clayton	12.09	1.72	7.0:1	0.68	68	6	26
70-072	Clayton	5.25	1.82	2.9:1	0.93	75	4	21
70-073	Clayton	7.80	3.58	2.2:1	0.51	82	9	9
70-074	Wilcox4/	29.22	4.40	6.6:1	1.04	6	37	57
70-075	Wilcox4/	29.89	7.19	4.2:1	0.08	90	4	6
70-076	Wilcox4/	11.63	3.07	3.8:1	1.04	26	48	26
70-077	Meridian	3.86	1.38	2.8:1	0.92	75	11	14
70-078	Porters Creek	10.17	3.04	3.3:1	1.73	37	55	8
70-079	Ackerman-Fearn Springs	7.39	1.61	4.6:1	0.16	85	5	10
70-080	" "	8.15	1.97	4.1:1	0.50	56	23	21
70-081	Porters Creek	12.30	1.60	7.7:1	2.34	0	57	43
70-082	Ackerman	7.96	2.51	3.2:1	1.80	72	17	11
70-083	Ackerman	10.99	3.76	2.9:1	2.43	49	40	11
70-084	Ackerman	7.65	2.10	3.6:1	0.31	56	30	14
70-085	Ackerman	7.49	1.26	5.9:1	0.26	77	15	8
70-086	Ackerman(lignite)	3.66	2.78	1.3:1	0.13	74	15	11
70-087	Ackerman	9.92	2.87	3.5:1	2.01	4	64	32
70-088	Ackerman	5.76	1.88	3.1:1	1.74	62	23	15
70-089	Ackerman	7.35	1.98	3.7:1	1.71	59	25	16
70-090	Ackerman	4.43	1.26	3.5:1	0.77	55	26	19
70-091	Ackerman	11.54	3.46	3.3:1	1.29	3	56	41
70-092	Ackerman(lignite)	9.09	8.73	1.0:1	0.32	13	28	59
70-093	Ackerman	11.20	10.87	1.0:1	1.51	1	49	50
70-094	Ackerman	13.41	3.50	3.8:1	1.22	3	57	40
70-095	Ackerman-Meridian	5.02	1.33	3.8:1	0.05	93	2	5

Appendix I: Thorium, Uranium and Potassium Concentrations and Particle Size Measured in Geologic Material from the Little Tallahatchie River Watershed, Mississippi (Continued).

Sample No.	Geologic Material	Thorium (ppm)	Uranium (ppm)	Th/U Ratio	Potassium %	Sand1/ %	Silt2/ %	Clay3/ %
70-096	Meridian	2.36	0.60	3.9:1	0.01	98	1	1
70-097	Meridian	2.57	0.49	5.2:1	0.06	94	2	4
70-098	Ackerman	7.55	1.46	5.2:1	1.40	58	24	18
70-099	Ackerman	9.48	2.09	4.5:1	0.92	27	41	32
70-100X	Meridian	2.66	1.53	1.7:1	0.17	85	4	11

1/ 0.05 - 2.00 mm
2/ 0.002 - 0.05 mm
3/ <0.002 mm
4/ Undifferentiated Wilcox samples.
5/ McNairy sand member of the Ripley formation.
6/ Not detected

CRETACEOUS - TERTIARY BOUNDARY OF EAST - CENTRAL GEORGIA AND WEST-CENTRAL SOUTH CAROLINA

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ABSTRACT

Previous recognition of nonmarine Cretaceous sediments of east-central Georgia and west-central South Carolina has been based on the occurrence of massive, white kaolin units and cross-bedded kaolinitic sands.

Pollen from the base of massive gray-white kaolin deposits indicates major kaolin deposition was not restricted to the Late Cretaceous, but also occurred during Middle(?) Eocene.

Opaline sponge spicules in the upper three feet of a highly-indurated (flint) kaolin unit of east-central Georgia suggests these sediments were not deposited during early Late Cretaceous, but Middle(?) Eocene.

Massive Eocene gray-white kaolin and flint kaolin deposits are separated from underlying Middendorf (Tuscaloosa) sediments by large scale cross-bedded kaolinitic sands. These sands contain large (up to one meter in diameter) kaolin clay balls that locally form clay ball conglomerates indicating extensive erosion of major portions of the underlying Middendorf clay deposits.

INTRODUCTION

Paucity of Cretaceous fossils has resulted in controversy regarding the Cretaceous-Tertiary boundary of east-central Georgia and west-central South Carolina. Throughout this area basal sedimentary rocks of the Coastal Plain rest nonconformably on deeply weathered

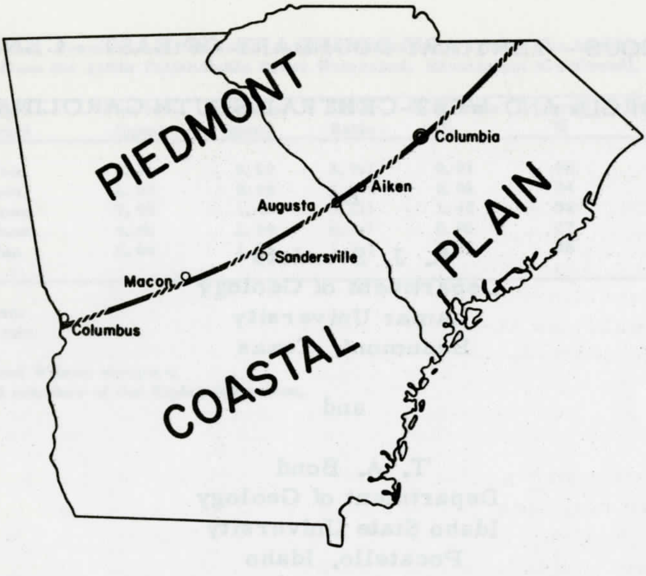

Figure 1. General location map of study area and Georgia-South Carolina Piedmont-Coastal Plain Contact.

pre-Cretaceous crystalline igneous and metamorphic rocks. These predominantly fluvial, lower Upper Cretaceous sediments strike about N55°E (Eargle, 1955); dip southeasterly at three to eight meters per kilometer and merge with marine equivalents of southern Georgia and northern Florida (Herrick and Vorhis, 1963). Basal Upper Cretaceous sediments thin to the northwest where they are in contact with the deeply-weathered igneous and metamorphic rocks of the Piedmont Province (Figure 1).

Major emphasis by previous investigators has been placed on the origin of massive, pure kaolin deposits interbedded with the cross-bedded kaolinitic sands of the Middendorf (Tuscaloosa) Formation (Toumey, 1848; Veatch, 1908; Neumann, 1927; Kesler, 1963; Jonas, 1964). Lower Eocene sediments have not been recognized in east-central Georgia or west-central South Carolina. These sediments either were not deposited or were eroded prior to deposition of Claiborne-equivalent units.

Middle Eocene sediments consist of cross-bedded kaolinitic sands, flint kaolin, massive kaolin units, spiculitic cherts and silicified biomicrudites. These sediments rest disconformably on basal Upper Cretaceous clastics throughout most of east-central Georgia and west-central South Carolina. Locally, and in close proximity to the Fall Line, these sediments rest nonconformably on pre-Cretaceous Piedmont rocks.

Along the Fall Line local variations in the thickness of Middle

Eocene sediments can be attributed to Middle Eocene and post-Middle Eocene erosion. Locally, Middle Eocene units have been completely removed and Upper Eocene or younger sediments rest unconformably on basal Upper Cretaceous or pre-Cretaceous metamorphic and igneous rocks of the Piedmont.

The Upper Eocene Barnwell Formation is composed predominantly of medium to coarse-grained ferruginous sands; thinly laminated, buff to light gray, fossiliferous clays; extensively silicified limestones and locally phosphatic cherts. Along the Fall Line, Barnwell sediments rest disconformably on Middendorf sands and clays and deeply weathered metamorphic and igneous crystallines of the Piedmont Province.

Previous Investigations

The outcropping sands, gravels and clays of the lower Upper Cretaceous have been designated as the Tuscaloosa Formation and Tuscaloosa Group (Veatch, 1908, 1909; Veatch and Stephenson, 1911, Neumann, 1927; Smith, 1929; Herrick, 1961; Herrick and Vorhis, 1963). Snipes (1965) separated lower Upper Cretaceous sands, clays and gravels between the Lynches River, South Carolina, and the Ocmulgee River, Georgia, from the Tuscaloosa Group and assigned these sediments to the Middendorf Formation. He based his conclusion on the similarity of these sediments to the type section of the Middendorf Formation exposed at Middendorf, South Carolina, and dissimilarity from typical Tuscaloosa sediments exposed at Tuscaloosa, Alabama. Based on field relations, Snipes' conclusions seem well-founded and basal Upper Cretaceous clastics are referred to as the Middendorf Formation in this article.

West of the Ocmulgee River Upper Cretaceous, predominantly fluvial, sediments grade laterally into marine limestones, fossiliferous shales, glauconitic sands, nonmarine carbonaceous shales, and quartzitic argillaceous sands.

Along the east-central Georgia Fall Line, lower Upper Cretaceous nonmarine clastic sediments of the Middendorf Formation rest nonconformably on crystalline rocks of the Piedmont. To the southeast these nonmarine clastics merge with the subsurface marine equivalents of southern Georgia and northern Florida (Herrick and Vorhis, 1963).

CRETACEOUS - TERTIARY DISCONFORMITY

Diagnostic of the Middendorf Formation are white, massive kaolin deposits and cross-bedded, kaolinitic, coarse-to-medium grained quartz sands and gravels. The massive kaolin deposits locally thicken to eighteen meters, but are commonly six to nine meters thick. Overlying and locally truncating these massive deposits are unconsolidated, poorly to moderately sorted, coarse-grained, micaceous, kaolinitic quartz sands. Throughout the Georgia and South Carolina "kaolin belt" these typically alluvial deposits can be seen truncating the massive white clay deposits.

In eastern east-central Georgia, a white to buff, highly indurated, massive clay unit overlies the cross-bedded, kaolinitic sands. This indurated clay has been locally referred to as flint clay or fire clay and has been assigned to the upper member of the Tuscaloosa (Middendorf) Formation (Veatch, 1909; Smith, 1929). The degree of induration changes rapidly within short lateral and vertical distances, causing the unit to vary from a friable, unconsolidated, arenaceous clay to a hard, flint-like rock that breaks with a conchoidal fracture.

The flint kaolin is extensively burrowed, evident in hand specimen by iron oxide linings along burrow walls. Cavities in the flint kaolin are occasionally filled with a colorless to light-blue gel. Allen (1952) noted that the silica gel was opaline in character, "being isotropic and having an index of refraction of 1.465."

Veatch (1909) and Smith (1929) believed the flint kaolin unit to be restricted to Glascock County, Georgia. This unit has been informally referred to as the Glascock Member of the Tuscaloosa Formation (Hurst, et al., 1966). Subsequent mapping has shown that the unit has a much wider distribution and occurs sporadically throughout eastern east-central Georgia.

Ths upper one meter of the highly-indurated massive clay contains opaline sponge spicules. These sponge spicules appear identical to spicules found in conformably overlying Middle Eocene carbonaceous spiculitic cherts suggesting deposition was continuous between the two units.

Deposition of the indurated clay was closely followed by a marine transgression and deposition of the carbonaceous spiculitic clay. A whale vertebra and well preserved plant fossils (Thrinax eocenica, Berry, 1914) from exposures at the Albion Kaolin Mine, Hephzibah, Georgia, indicate that this unit was deposited in a near-shore environment.

In western east-central Georgia light-to-dark gray massive kaolin deposits directly overlie the cross-bedded, kaolinitic sands. Locally, these clays thicken to thirty meters. The gray color is related to carbonaceous plant remains which locally forms up to seventy percent of the unit. The higher organic content occurs at the base of the unit.

A palynological assemblage was recovered from an argillaceous lignite at the base of a twelve meter section of clay, southwestern Columbia County, Georgia, and from the base of a one meter bed of lignitic clay approximately twenty kilometers west of Sandersville, Georgia.

A total of seventeen different spore and pollen types were found in the samples. The pollen spectra made up approximately ninety-nine percent of the total and the only two spore types recognized were assigned to the genera Cyathidites and Cicatricosporites. The forms assignable to these two genera were poorly preserved and have probably been reworked from Cretaceous sediments. It was impossible to establish affinities for most of the pollen, but in many cases, either a positive identification or an identification based on suspected relationships was made.

The large number of microfossils would indicate either a tropical to subtropical flora, or a climax condition with many individuals but few genera represented. The type of fossils found seem to bear out the former. Monosulcate forms believed to be related to the palms were found as well as several cycad-like grains. This is to be expected as the spiculitic clays contain well preserved macrofossils of Thrinax eocenica (Berry, 1914) which is assigned to the Family Arecaceae (Palmae). The most abundant pollen types found were Englehardtia, Cyrilla, Ilex, Taxodium, Momipites and two distinct types of Tricolpites. Englehardtia was the most abundant making up almost thirty-five percent of the total pollen count. Several grains tentatively assigned to the genus Castanea were also recorded. Although Castanea is usually regarded as being a temperate genus, there is a species now living in Florida. The sparse presence of anamophilous Castanea pollen is probably indicative of an Eocene upland forest some distance from the depositional site of the lignite.

The lignite itself is probably a fresh-water swamp deposit. Taxodium was found and this genus is known to exist in a swampy semi-aquatic environment. The lignite was laid down in a sub-tropical climate and the fresh water swamp was probably some distance from a land area as there is a distinct lack of upland forest pollen types. It is also noteworthy that there is no gymnospernous pollen recorded and only a trace of spores which could be assigned to the lower plants.

Stratigraphically the massive gray kaolins of western east-central Georgia appear correlative with the highly-indurated kaolin units of eastern east-central Georgia (Figure 2). These massive gray kaolin deposits are separated from the Middendorf Formation by massive, cross-bedded, kaolinitic sands making an erosional disconformity separating Cretaceous from Eocene units. The cross-bedded kaolinitic sands have locally eroded underlying kaolin deposits evidenced by cut and fill structures in the kaolin deposits and by large, up to two feet in diameter, kaolin balls found throughout the cross-bedded kaolinitic sands.

CONCLUSIONS

Recognition of Middendorf (Tuscaloosa) sediments of east-central Georgia and west-central South Carolina should not be based on the presence of cross-bedded kaolinitic sands and massive kaolin units. Pollen indicates two major periods of kaolin deposition occured throughout this South Carolina and Georgia "kaolin belt."

Basal Upper Cretaceous massive clays are separated from

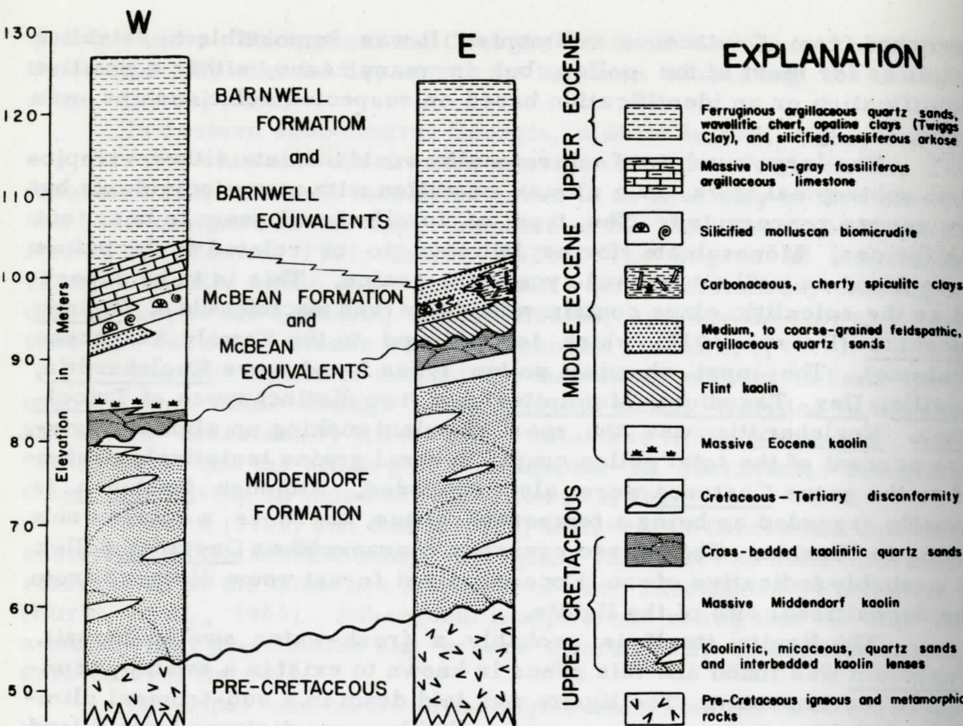

Figure 2. Stratigraphic sections of the eastern (Augusta) and western portion (Sandersville) of the area studied.

Eocene clays by large scale, cross-bedded, kaolinitic sands. Locally, these cross-bedded units can be seen truncating the massive Middendorf clay units in which large portions of the Middendorf clays have been removed by erosion.

These typically alluvial, cross-bedded kaolinitic sands are representative of a major erosional disconformity separating highly indurated flint kaolins from the massive Middendorf units in eastern east-central Georgia and west-central South Carolina. Massive, cross-bedded sands separate basal Upper Cretaceous units from massive gray kaolin units of east-central Georgia.

Stratigraphically, the flint kaolin units are equivalent to the massive, gray-white, Eocene kaolin deposits. Eocene deposition for the flint kaolin is further demonstrated by the presence of opaline sponge spicules in the upper section.

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MISSISSIPPI GULF COAST PLEISTOCENE BEACH BARRIERS AND THE AGE PROBLEM OF THE ATLANTIC-GULF COAST "PAMLICO" -- "INGLESIDE" BEACH RIDGE SYSTEM

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ABSTRACT

The detailed stratigraphic and geomorphological study of the Pleistocene coastal Mississippi barrier ridge and seven absolute age dates from the complex and areas adjoining it add to the accumulating body of evidence that the discontinuous Ingleside-Pamlico mainland beach ridge-barrier island trend along the Gulf of Mexico and Atlantic shores is of Sangamon interglacial origin. "Dead" radiocarbon dates place the time of formation before radiocarbon dated Late Pleistocene interstadials.

During the transgression that culminated in the formation of the barrier ridges, sea level rose by at least 15 m, followed by a maximum 4.5 km beach ridge progradation. This had to be the result of an extended (interglacial) warm period with greater deglaciation than at the present time. All the known Wisconsin interstadials were too brief to account for the dimensions and the associated high sea level elevations of this barrier system. The Late Pleistocene (Sangamon) interglacial sea level stood about 7 m above the present sea level along the Mississippi Coast. No surface or subsurface evidence has yet been found for the existence of any higher level, Pre-Sangamon marine Pleistocene littoral ridge system in the southern Mississippi-Alabama and in the Florida Panhandle coastal zone.

INTRODUCTION

The stratigraphic investigation of the Mississippi coastal barrier ridge area, including seven"dead" radiocarbon and one Th/U date sheds additional light on the age of one of the major Pleistocene littoral barrier ridge systems. This discontinuous chain of fossil mainland beaches and barrier islands girdles most of our south Atlantic and Gulf of Mexico coasts and is the most extensive Late Neogene coastal barrier complex along these coasts, save the Recent one (Fig. 1; "Early Sangamon" Atlantic coastal plain, Talbot - Suffolk Scarp-barrier ridges are not shown on map). This barrier complex is assumed to be time-correlative with one of the ridge systems, the surfaces of which constitutes a "staircase" of coastwise plains ("terraces") along the south Atlantic coast, extensively described by C. W. Cooke between 1924-45 (Cooke, 1966, p. 39-40; List of References). Six of these surfaces (Wicomico, Penholoway, Talbot, Pamlico, Princess Anne and Silver Bluff), ranging in elevation between 37.5-1.5 m above present sea level, are generally considered Pleistocene and at least partly coastal marine in origin. Each has a characteristic elevation range of its own (Richards, 1969). Cooke's nomenclature of these surfaces was also applied by several workers to parts of the northern Gulf of Mexico coastal area. Physiographic characteristics (elevation values, surface topographical features) have often been used in such age determinations and surface-correlations, frequently to the total exclusion of sedimentological and paleontological analyses and without regard to the actual, often non-marine depositional environments of the sediments underlying the surfaces. Recent papers (Oaks and Coch, 1963; Hoyt and Hails, 1967; Colquhoun et al., 1968 and Colquhoun, 1969a; 1969b) established the beach ridge, barrier island and lagoonal marsh origins of deposits which underly many of the Atlantic Pleistocene coastal surfaces ("terraces"). A single discontinuous chain of Pleistocene barrier ridges was also identified in the Texas-southwest Louisiana coast (Ingleside-Live Oak Chain; Price, 1933; Bernard and LeBlanc, 1965 and Graf, 1966) and in certain parts of the northern-northeastern Gulf of Mexico coast.

Surface elevations of adjoining lagoonal salt marsh sediments are now being utilized as reference elevations of corresponding sea level stands along barrier ridges (Graf, 1966; Hoyt and Hails, 1967). This eliminates possible errors of measuring ancient sea level elevations on the top of fossil dunes which at the time of ridge development may have extended high above their contemporary sea levels. Because of post-depositional tectonism, these ancient sea level values, especially in the case of older barrier systems, do not necessarily indicate sea level stands everywhere, close to these present day reference elevations. In southeast Texas - southwest Louisiana the maximum 7 m elevation-differences between ancient lagoonal surfaces behind members of the Ingleside barrier chain (Graf, 1966) are related to subsidence south of the hinge line down to almost the present sea level. Uplift of the Florida Peninsular Arch was also reflected in significant differences between present day barrier elevations of various ridge units in northeastern Florida and eastern Georgia (Hoyt, 1969).

In contrast with the northern Gulf shores, a number of Pleistocene barrier ridge systems occupy the south-Atlantic Coastal Plain. The most extensive of them, in the southeast Atlantic states is a ridge set of Pamlico age, with lagoonal surface elevations usually 6-7 m above the present sea level. At most locations it is well-defined in relation to the older, higher Talbot (9.5-15 m) and the younger, lower, localized Princess Anne (2.7-6 m) ridges (Richards, 1969; Colquhoun, 1969). The Pamlico ridge trend has good linear continuity between the North-South Carolina border region and east-central Florida. Elevations of the Gulf Coast barrier ridge chain, including its sections in northwestern peninsular Florida close to the Atlantic Pamlico ridges (Vernon, 1951) are compatible with those of the Pamlico. No beach ridges were preserved between the Gulf and Atlantic barrier ridges in Western and in south-central peninsular Florida (Cooke, 1945, Figure 1). Despite this lack of perfect physical continuity, the absence in the barriers of correlatable fossil horizons of time-stratigraphic value, and despite the lack of finite absolute dates, the time-correlative nature of the Atlantic and Gulf ridge segments seems highly likely because of their compatible elevations that appear to be related to the last interglacial high sea level stand, documented by absolute dates (Broecker and Thurber, 1965) in the tectonically stable south Florida region.

Recent investigations (Otvos, 1972) failed to confirm the presence of any alleged Talbot, Penholoway, Wicomico, Sunderland and other high, earlier Pleistocene littoral ridges ("marine bars" and "terraces") along the northeastern (Mississippi, Alabama and Florida Panhandle) coast of the Gulf of Mexico. No such littoral ridges are known to have been described from the northern and northwestern Gulf coastal areas. Figure 2, accordingly, shows none of these alleged features that appear on maps in several publications and reports of the subject areas. Fluvial erosion during low sea level glacial stages and more intense earlier Pleistocene regional tectonic activity (epeirogenic uplift) in the Gulf region than in the Atlantic seaboard area may account for this.

Acknowledgments

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RADIOCARBON AGE DATING AND THE AGE OF THE PAMLICO-INGLESIDE ATLANTIC-GULF COAST BARRIER SYSTEMS

Although the interglacial origin of all the Pleistocene barrier system is generally accepted, opinions are still strongly divided and no concensus has been reached on the Mid-Wisconsin "interglacial" (Cooke, 1945 and numerous other authors) or Sangamon interglacial (Colquhoun, 1969a, and many others) of the Pamlico. A Mid-Wisconsin "interglacial"

(Bernard and LeBlanc, 1965; Graf, 1966) or Sangamon age (Richards, 1939) was assumed for the Ingleside-Live Oak barrier system and for related Beaumont - Prairie sediments. This "Mid-Wisconsin" warm period was suggested for 85,000-60,000 yrs. B. P. (Bernard and Le-Blanc, 1965) and 60,000-50,000 yrs. B. P. (Graf, 1966). At present an 80,000-70,000 yrs. B. P. date is accepted by many for the end of the Sangamon interglacial and the beginning of the Wisconsin glacial.

Radiocarbon dates from the Ingleside-Pamlico barrier system, as well as from Wisconsin deposits in general, do not support the existence of a Mid-Wisconsin extended warm interglacial period and corresponding high sea level stand. Finite Mid-Wisconsin dates from a barrier sequence in coastal North Carolina, if not influenced by radiocarbon contamination (Whitehead and Doyle, 1969), reflect the age of an upper dune sand and fresh water peat-bearing section in the barrier ridge. There are, on the other hand, several "dead" dates related to this barrier system. Radiocarbon dates, with values from 35,000+ to 40,200+ yrs. B.P. were derived from Delaware (Kraft, 1971), North Carolina (Oaks, 1965, p. 169), South Carolina (DuBar, 1962), southwestern Louisiana (Graf, 1966), Texas (Pampe, 1971) and southern Mississippi (present paper) from Pamlico-Ingleside barrier complex deposits.

In addition to these dates, 42,000+ B. P. ("dead") and 24,000-40,000 yrs. B. P. finite dates were published from the younger Princess Anne and probable Silver Bluff sediments in Georgia (Hoyt et al., 1968) and in the Florida Panhandle (Schnable and Goodell, 1968), respectively. These suggest that the Pamlico ridges must be considerably older than Mid-Wisconsin and that in Georgia during the Mid-Late Wisconsin the sea level was less than 11-12 m below present sea level but no proof was given to show that it was any higher. To date no conclusive finite Mid-Wisconsin dates are known to the present writer from beach and shoreface facies of the Ingleside-Pamlico or Princess Anne ridges.

MISSISSIPPI COAST PLEISTOCENE BARRIERS AND THE AGE PROBLEM

Absolute dates from the Mississippi coastal barrier complex area suggest additional confirmation of the Sangamon age of these barriers. There are three ridge complex segments on the Mississippi Coast between Waveland and Belle Fontaine Point. Their widths range between 1.5-4 km and the cumulative length is about 57 km (Figure 2). The existence of the Waveland - Bay St. Louis segment has not been noted previously. Deer and Round Islands in the Mississippi Sound are related to the Belle Fontaine segment, as their thin Recent dune-beach sand veneers are underlain by well-sorted, better consolidated, humate-bearing, dark gray and oxidized yellowish brown Pleistocene, occasionally silty, sand units. A humate-impregnated sand sample from northwest Deer Island (intertidal zone) yielded a 16,635 ± 255 yrs. B.P. date (Univ. of Georgia Geochronology Laboratory; No. UGa-385; 8-1/3-1972), owing to possibly Holocene contamination of the original Sangamon sediments.

The fine-grained, well-sorted, occasionally silty barrier sands often show no macroscopic sedimentary structures at all, although at several locations they do exhibit horizontal lamination. Humate impregnations were found in numerous outcrops and borings. Probably due to postdepositional leaching, the sands contain no microfauna. Macrofossils were found only at the cited Bay St. Louis outcrop. Current stratigraphic core drilling by the Geology Division, Gulf Coast Research Laboratory, has shown that the barrier sand unit (Gulfport Formation; Otvos, 1972 and in prep.) is maximum 8-14 m thick and downward usually grades into a sandy, silty clayey, maximum 5-13 m thick transgressive unit (Biloxi Formation; Otvos, 1972 and in prep.) which frequently contains rich molluscan and foraminifera faunas.

The maximum ridge elevations of the coastal barrier complex range between 3-10 above sea level. At places as many as 6-7 parallel coastwise ridges are present with corresponding swales. The ridge sequence represents local regression of the Sangamon sea, the level of which remained stable during barrier outbuilding. No lagoonal and barrier island units were yet recognized- the barrier complex flanked the mainland shore.

On the western and eastern flanks of the present Mississippi Coast larger streams entered the sea, depositing alluvial-deltaic "Prairie" sediments along the shore and thus preventing barrier ridge formation. Most of the present surface boundaries of the barrier segments are erosional and characterized by ridge slopes significantly steeper (2^o-8^o) than slopes of the more centrally located ridges. The steeper slopes along the northern margin of the barrier complex were created dominatly by Late Pleistocene and earlier Holocene fluvial erosion, while slopes on the southern margin were steepened by Late Holocene marine erosion along the Mississippi Sound.

Contrary to certain statements (Bernard and LeBlanc, 1965, p. 150), no drastic tectonic changes, similar to the ones found in the Ingleside chain near Galveston Bay, can be detected to have influenced the position of these barrier ridges. Judging from the presence of marine shells and beach foreshore-type lamination in Bay St. Louis, the ancient sea level was about 7 m above the present sea level and not as low (as has been suggested) as the beach levels of the present Mississippi Sound barrier islands. Identical foreshore - type beach lamination was found in a big Gulfport sandpit (By Silver Ridge Ave.) at 4-7 m above present sea level.

Seven "dead" radiocarbon dates were received from the Gulfport Formation and from associated units. Prairie (=Pamlico) age alluvial deposits, including partially festoon-bedded river channel sands in a large borrow pit included peat beds of probable backswamp origin about 2 m above sea level and 4 m below the Prairie land surface (Location 2, Figure 2; SW 1/4 of SE 1/4 of sec. 21, T. 7 S., R. 12 W). The peat's date: 41,000+ yrs. B. P. ("dead"; Institute of Marine Sciences, University of Miami; No. ML-680; 11-10-1970). The Prairie alluvium was deposited landward of the coastal barrier complex, contemporaneously with its development. Humate impregnation of the barrier sand at Belle Fontaine Point (Location 3, Figure 2) has been dated 35,000+ yrs. B. P. ("dead"; written commun. by W. S. Moore, U. S. Naval Oceanographic Office, Washington, D. C.). Humate-impregnated sand from sea level elevations in Biloxi (corner of Rodenberg Ave. and Highway 90; Location 6) yielded a 40,000+ yrs. B. P. date (University of California, San Diego, Radiocarbon Laboratory; No. L. J. 2471; 5-12-1972) and shell fragments from the Late Pleistocene Biloxi Formation (Otvos, in prep.) in Gulfport (corner of Lewis Ave. and Highway 90; Location 5) at 6 m below sea level were dated 40,000+ yrs. B.P. (same Laboratory; No. L.J. 2470; 4-25-1972). Peat beds from a fine-grained-sandy Prairie alluvial sequence north of Ocean Springs (Van Cleave Rd.; Location 7) gave a 27,000+ yrs. B. P. date (Krueger Enterprises Inc., No. GX2654; 5-11-1972). Upright-standing lignitized tree trunks near sea level elevations are covered and surrounded by the oyster-bearing estuarine silty clay facies of the Biloxi Formation in the southern bank of the Industrial Seaway, Gulfport (Lorraine Road bridge; Location 4). The trunks also, were dated "dead" (42,000+ yrs. B. P., University of California, San Diego, Radiocarbon Laboratory; No. L. J. 2294; 11-1-1971).

Oyster fragments from the Biloxi Formation in the Pass Christian City Park core well yielded a incongruous 37,100 ± 1800 yrs. B. P. date (Location 3; 18-26 ft.; Univ. of Georgia Geochronology Laboratory; No. UGa-370; 7-6-1972). The level of measured radiation was less than 0.2 counts per minute above background. There is no doubt that the actual age is much greater and postdepositional contamination was responsible for the false finite date. False finite dates, caused probably by percolating solutions of humic acids, were derived from wood fragments of pre-Holocene southeastern Louisiana fluvial deposits. A similar false date also came from woody materials in the Biloxi Formation at Location 4 (25,350 ± 110 yrs. B. P.; Geochron Laboratories; No. GX-2106; 12-9-1970).

Marine shells were found in a scarp created by the 1969 hurricane in Bay St. Louis beneath North Beach Blvd., near the intersection of DeMontluzin Ave. at 4.4-6.0 m above sea level (Figure 2, Location 1). Parallel-laminated layers of well sorted fine sand, apparently deposited in a foreshore environment, contained well-preserved Crassostrea virginica and Rangia cuneata specimens, typical of coastal facies. Three samples were subsequently dated with the Th/U method by J. K. Osmond at Florida State University. Only one of the samples (Crassostrea; No. MC2B) had a relatively lower U 234/U 238 ratio, indicating that the shell may have remained a closed system:

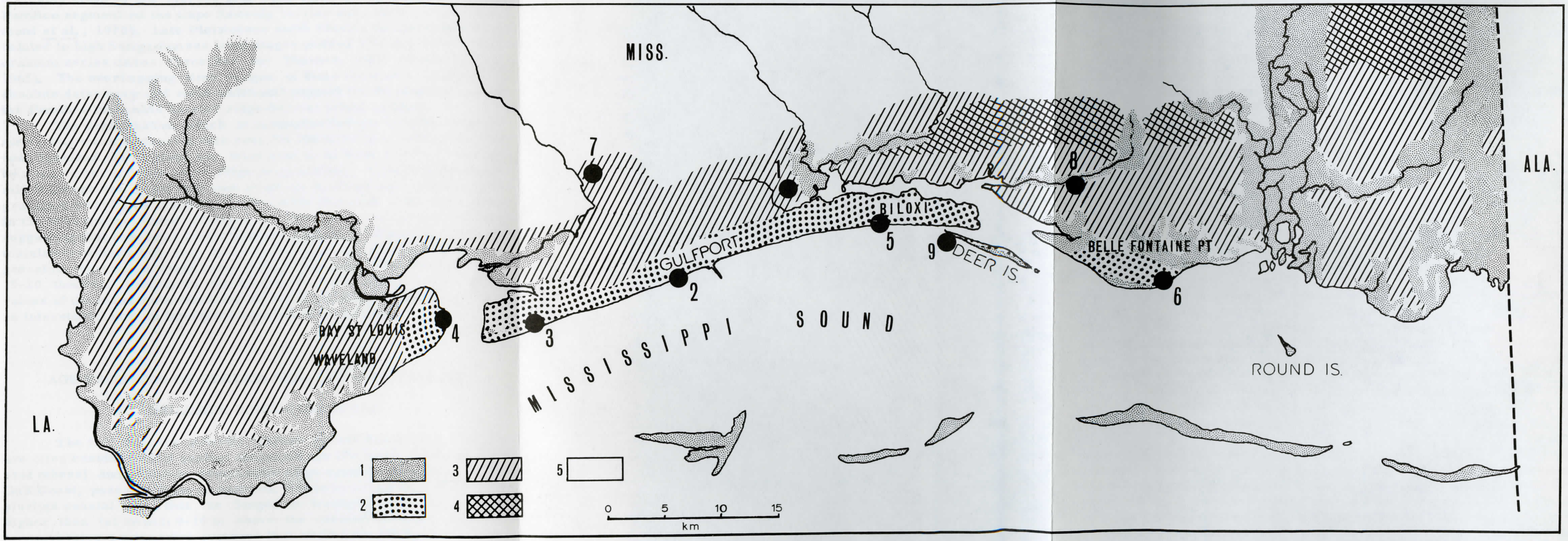

Figure 2. Mississippi Coast Barrier Beach System.

U ppm	Th ppm	U 234/238	Th 232/U 234	Th 230/U 234	Apparent Age (B. P.)
		Alpha - activity ratios			
.32	.22	1.45	.12	.59	95,000 years

Despite the fact that this one is the sole "reasonable" Th/U date and that the reliability of the Th/U method for brackish water shells has been strongly questioned recently (written comm. by Wm. Sackett; A. Kaufman et al., 1971), it still may be of some value. The age is remarkably close to the 110,000 yrs. Th/U date, derived from sites in the Pamlico segment of the Cape Kennedy barrier and lagoon complex (Osmond et al., 1970). Late Pleistocene south Florida marine limestones, related to high Sangamon sea level stages yielded 100-170,000 yrs. B.P. uranium series dates (Broecker and Thurber, 1965; Osmond et al., 1965). The overlapping time ranges of these three sets of consistent absolute dates may add some additional support for the Sangamon age of the Pamlico-Ingleside barrier ridge-barrier island systems.

The deglaciation which accompanied barrier complex development, was more extensive than even the Holocene one. Within the present mainland coast alone sea level rose by at least 15-17 m, followed by a maximum 4.5 km beach ridge progradation. A comparable early-mid Holocene sea level rise took about six thousand years and ridge progradation must have taken an additional few thousand. Drill cores taken in the Mississippi Sound (Rainwater, 1964) seem to contain evidence suggesting that the total Late Pleistocene sea level rise in the Sound and mainland coastal area may have exceeded 30 m. Clearly, this transgression-regression cycle in the coastal zone must have lasted at least 15-20 thousand years; in all likelihood much longer. The elevation values of the maximum sea level and the length of the cycle suggest not an interstadial but an interglacial period of formation.

AGE PROBLEM OF CERTAIN PRE- AND POST-PAMLICO-INGLESIDE RIDGE COMPLEXES

The fact that the Atlantic Talbot- Suffolk Scarp barrier ridges are often considered of earlier Sangamon age (Richards, 1969 and several others) and that no comparable ridge system can be found on the Gulf Coast, poses a difficult problem. There is no evidence in the Mississippi coastal area that the Sangamon transgression ever reached higher than (at most) 9-10 m above the present sea level. Another transgressive-regressive sequence that would have reached 3-6 m higher than this elevation during the same interglacial, just a few tens of thousands years earlier, would definitely have left more than a trace in the geological record. A pre-Illinoisan age of the Talbot and associated ridge systems would provide a longer time span during which geological processes might have prevented formation of such ridges or (more likely) would have eliminated ridges already formed. Alternative conditions, as yet unsupported, would be the start of the Sangamon stage much earlier than known presently (170,000 yrs. B.P.; Broecker and Thurber, 1965) and/or extended colder, low sea level periods during the Sangamon.

In the Virginia coastal plain two regressions took place between the formation of the Suffolk Scarp (Talbot?) and the Oceana (Pamlico?) Ridges (Oaks, 1965) and these regressions were accompanied by considerable subaerial weathering and erosion. This certainly implies that substantial time must have elapsed between the formation times of the two ridge systems.

The post-Pamlico Princess Anne ridge set, on the other hand, could have formed during the later stages of the Late Sangamon-Early Wisconsin marine regression phase which began after the Pamlico-Ingleside barrier development ended. Falling Sangamon sea level brought Beaumont (Prairie) alluvial deposits south (sea-)ward of the Ingleside chain, completely surrounding the toes of these ridges (Bernard and LeBlanc; 1965, p. 142, Figure 5). About 50 km south of the Ingleside system, east of Orange, Texas, a new ridge system developed in association with this falling sea level. These ridges are as high (3.0-4.5 m) as members of the Atlantic Princess Anne complex and suggest development at the same time. No need or conclusive evidence exists as yet for postulating a much later, separate Princess Anne transgression-regression cycle ("submergence"; Hoyt et al., 1968, p. 391, Figure 2) forty-fifty thousand years ago.

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